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ELECTRIC AND THERMAL PROPERTIES OF ROCKS

by U. I. Moiseyenko, L. S. Sokolova,
and V. Ye. Istomin

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Siberian Department, Novosibirsk, 1970

NATIONAL AERONAUTICS AND SPACE ADMINISTRATION

ANNOTATION

This monograph gives the results of study of the electric conductivity of rocks under different thermodynamic conditions. The apparatus and methods used in measuring these parameters are described. Experiments for studying the effect of temperature, pressure, and the joint effect of these factors on rock resistivity are discussed, as well as the results of a study of the thermal properties of sedimentary and igneous rocks at temperatures from room temperature to 1,200-1,400°.

The dependence of electric conductivity is given for unilateral pressures and with simultaneous heating to 600°C and pressures greater than 30 kbar.

The results of a study of the dependence of heat conductivity on temperature are discussed.

Conclusions are drawn concerning the mechanism of electric and heat conductivity of rocks at high temperatures.

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FOREWORD

Investigations of the electric and thermal properties of rocks under different thermodynamic conditions, at high temperatures and pressures at which these parameters have been least studied, have now been made for several years in the Physics of the Earth's Crust Laboratory at the Institute of Geology and Geophysics, Siberian Department, USSR Academy of Sciences. A knowledge of the nature of their changes and their dependence on prevailing temperatures is extremely important for understanding many geological phenomena, as well as the nature of geophysical discontinuities in the earth's crust and the temperature distribution in its cross section.

The first stage in the investigations provided for formulating a method for measuring the effect of different factors on the physical properties of rocks. For example, in studying the resistivity of most igneous rocks we separately investigated the effect of temperature at atmospheric pressure, the effect of pressure at room temperature, and the joint effect of temperature and pressure.

In view of the paucity of information on the thermophysical properties of rocks in general, it is obviously of interest to study them in a broad range of changes in temperature and pressure, including at normal pressure and at room temperature. As yet we have studied only the dependence of the heat conductivity coefficient on temperature in the range from 20 to 1,200 - 1,400° C.

The collected experimental data on the electric and heat conductivity of rocks under different thermodynamic conditions made it possible to undertake an explanation of the mechanism of these phenomena.

In the investigations described in this monograph the authors were assisted by M. A. Aliyeva, A.A. Solov'yeva and V.A. Kutolin.

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ELECTRIC AND THERMAL PROPERTIES OF ROCKS

U. I. Moiseyenko, L. S. Sokolova and V. Ye. Istomin

INTRODUCTION

A considerable number of studies have now been published on the effect of temperature and pressure on the electric conductivity of rocks. However, far fewer studies have been published on the thermal properties of rocks.

Several years ago such information was published in individual collections of articles, such as F. Birch, H. Spicer and I. Scherer (1949) and V. N. Dakhnov and D. I. D'yakonov (1952). Accordingly, in this review, emphasis is on studies pertaining to study of the electric conductivity of rocks. Studies devoted to the thermal properties of these rocks, being so numerous, are mentioned only in the text of the corresponding section.

The studies of interest for our investigations of the electric conductivity of rocks can be divided into three groups: the first pertain to the dependence of electric conductivity only on temperature at normal atmospheric pressure, the second pertain only to the dependence on pressure at room temperature, and the third pertain to change in electric conductivity under the influence of varying temperature and pressure.

An increase in electric conductivity of rocks with a temperature increase was observed in all studies in the first group. Judging from the studies of different authors, the nature of the increase is somewhat different. Up to definite limits of temperature change (600-800°) this dependence is linear; for example, this was observed by V. A. Marinin (1938) and H. N. Coster (1948) for the electric conductivity of granites and gneisses at temperatures up to 750°, and E. I. Parkhomenko and A. T. Bondarenko (1962) for the electric conductivity of diabases, basalts, and peridotites in the range from 100 to 500-800°. At higher temperatures the linear dependence no longer prevails and in-

*Numbers in the margin indicate pagination in the foreign text.

flections appear on the curves. For example, H. N. Coster (Coster, 1948) ob-
served this for gabbro, basalt, peridotite and eclogite, and E. I. Parkhomenko
and A. T. Bondarenko (1962; Bondarenko, 1966) for diabase, pyroxenite, oliv-
inite, peridotite and andesitic basalt. The same thing was noted by Ye. B.
Lebedev and N. I. Khitarov (1964) in studying the electric conductivity of
granites.

K. Noritomi and A. Asada (Noritomi, Asada, 1956) note a somewhat different
nature of the dependence of electric conductivity on temperature for acidic and
intermediate rocks and also for quartz and perthite. In the temperature var-
iation of such curves the first inflection is observed at 450 to 500°. For
some serpentinite samples K. Noritomi observed an electric conductivity mini-
mum at a temperature of 600 to 700° with a subsequent increase. The electric
conductivity decreased monotonically with cooling of the sample. Other authors
have also observed such a change in electric conductivity with temperature,
that is, the presence of local minima on the dependence curve with its general
increase in the first cycle of sample heating and smoothing of such curves with
cooling and repeated heating. T. Murase (Murase, 1962) established this fact
for basalt, a number of lavas and obsidian. U. I. Moiseyenko and V. Ye.
Istomin (1963) did the same for dunite, pyroxenite, olivinite and granite, and
Yu. I. Protasov (1964) did the same for pyroxenite, granite, secondary quartz-
ite, and tuff-diabase. Since these authors studied rocks of different com-
position, it is difficult to compare the anomalous intervals on the curves
which they constructed.

Investigations of the effect of pressure on the resistivity of igneous
rocks have been made primarily in the Soviet Union.

E. I. Parkhomenko and A. T. Bondarenko (1960) carried out a series of
measurements of resistivity of igneous and sedimentary rocks under unilateral
pressures from 10 to 600 kg/cm^2 at room temperature with a dc current by the
guard ring method. The samples were cut in the form of disks 0.5 to 2.0 cm
high and 2.8 to 7.0 cm in diameter. In the course of increasing pressure in
all the investigated rocks there was a decrease in resistivity, but to a dif-
ferent degree. For some rocks these changes were 10 to 20% or more; for
others it was only a few percent. The maximum changes corresponded to the

range of unilateral pressures from 10 to 300 kg/cm^2. At higher mechanical pressures the resistivity changed insignificantly.

M. P. Volarovich and A. T. Bondarenko (1963) investigated resistivity in rock samples at a hydrostatic pressure up to 1,000 kg/cm^2; these revealed that the dependence observed in the case of unilateral pressure for the most part persists. Samples of basalt, peridotite, schist, and sandstone were investigated.

Under hydrostatic pressure resistivity changes more than under the influence of unilateral pressure; in the first case by 20 to 40%, and in the second case by 5 to 20%.

U. I. Moiseyenko, V. Ye. Istomin and G. D. Ushakov (1964) increased the pressure on the sample considerably (unilateral to 20,000 kg/cm^2, hydrostatic to 2,000 to 3,000 kg/cm^2. Experiments made on samples of olivinite, marble, serpentinite, dunite, pyroxenite, basalt and peridotite yielded substantially new data. With a pressure increase the resistivity for all the investigated rocks decreases to a definite limit (corresponding to different pressures for different rocks). A further pressure increase leads to an increase in rock resistivity.

Studies in the third group are of the greatest interest in studying the behavior of resistivity when rock samples are heated at high pressures.

One of the first studies of this type was written by N. Hughes (Hughes, 1955). He studied the resistivity of peridotite under a hydrostatic pressure up to 10,000 kg/cm^2 at temperatures of 1,063, 1,143 and 1,210°. The direct and inverse resistivities were measured at each of these temperatures and at pressures of 1,000, 2,500, 4,000, 5,560, 7,000 and 8,500 kg/cm^2. N. Hughes mentions a decrease in the electric conductivity of peridotite with pressure (by 2.3 to 3.7% per each 1,000 kg/cm^2) at a constant temperature and an increase in electric conductivity with a temperature increase.

M. P. Volarovich, E. I. Parkhomenko and A. T. Bondarenko (1963, 1966) studied the resistivity of a number of rocks at a pressure greater than 30,000 kg/cm^2 and temperatures up to 400 to 600°. In the entire range of used pressures the authors detected two different types of dependence of resistivity on pressure. Some rocks are characterized by a continuous decrease in resistivity

with pressure; for others there is first a decrease in resistivity and then an increase in resistivity at higher pressures. The authors note that an increase in temperature to 300 to 400° exerts a considerably greater effect on electric conductivity than a pressure of 10,000 atmospheres. A petrographic study of sections of rock samples after the experiments revealed a structural change. E. I. Parkhomenko and A. T. Bondarenko (1963) indicate a fragmentation and formation of grains of some minerals, twinning in pyroxene grains, and the formation of deformation borders in serpentinized dunites. The authors do not feel that these changes are particularly important; however, we feel that the onset of an increase in resistivity with a pressure increase, especially on the curves for serpentinized dunite, is associated with the time of rock fragmentation.

The behavior of the resistivity of olivinite, serpentinite and eclogite was studied by U. I. Moiseyenko and V. Ye. Istomin (1964) at a temperature of 600° and a pressure up to 30,000 kg/cm^2. Under these conditions a constant decrease in resistivity was observed with a pressure increase.

Ye. B. Lebedev and N. I. Khitarov (1964) observed a considerable decrease in the resistivity of granites (in the presence of water) in a pressure range up to 9,000 kg/cm^2 and at temperatures from 600 to 1,200°. A communication by R. S. Bradley, A. K. Tamil and D. S. Munro (Bradley, Tamil, Munro, 1962) also mentions a decrease in the resistivity of fayalite and spinel with a pressure increase to 35,000 kg/cm^2 at temperatures up to 680°.

Most researchers engaged in a study of the behavior of electric conductivity at high temperatures endeavor to clarify the electric conductivity mechanism. The activation energy E and the value of the preexponential term σ_0 for the derived dependences were computed for this purpose. E. I. Parkhomenko (1965) published summarized σ_0 and E data for rocks of acidic, intermediate, basic and ultrabasic composition. She concludes on the basis of data from a number of researchers that the activation energy E for intermediate and acidic rocks is low, 0.7 to 0.9 eV for temperatures below 600 to 700°. With an increase in temperature the activation energy increases and in the range 1,000 to 1,200° attains 4 to 12.5 eV. The activation energy of basic and ultrabasic rocks at a temperature of 650° is 0.6 to 0.9 eV, which in some cases is close to the activation energy of acidic and intermediate rocks. When T > 800° E increases to 1.6 eV and only in some cases has higher values. At high temperatures (for

example, for granite more than 1,250° and for diabase 870°) the author notes a decrease in activation energy, relating it to the melting of the rock.

Special experiments for clarifying the nature of the current carriers were carried out by N. Coster (Coster, 1948). The author concludes that electrons as well as ions participate in the current transfer. K. Noritomi, et al., (1955, 1956), in analyzing their experiments and the results of earlier studies, concluded that in olivines or rocks having the structure of olivine, the electric conductivity at $T \approx 600°$ corresponds to so-called extrinsic conductivity, in the range 600 to 1100° to semiconductor in combination with ionic conductivity, and at $T = 1100°$ to ionic conductivity. The comparison of the chemical composition of acidic and intermediate rocks and their activation energies made by K. Noritomi (1961) enables him to postulate a substantial role of SiO_2 and Al_2O_3 compounds in the mechanism of rock electric conductivity. T. Murase (Murase, 1962) also notes some dependence of activation energy on the compound SiO_2.

F. S. Zakirova (1964) postulates that in the region 700 to 1500°, that is, between the first and second inflections on the dependence curves, the current carriers are potassium and sodium ions. At high temperatures bivalent ions participate in current transfer. In addition, for rocks with an identical potassium content, F. S. Zakirova notes a relationship between electric conductivity at the second curve inflection point and rock age. The greater the age of the rock, the lesser is its σ; this is related to an accumulation of valency, occurring as a result of the radioactive decay of K^{40} in which Ca^{40} is formed.

A detailed study of the electric conductivity mechanism was made by R. Hamilton (Hamilton, 1965); he investigated the temperature, composition and conductivity mechanism for the upper mantle. For olivines of different composition he noted a decrease in activation energy with a pressure increase and a dependence of conductivity on fayalite content. On the basis of an analysis of studies on the electric conductivity of minerals possibly constituting the upper mantle, the author concludes that clarification of the type of charge carriers is the most important problem in studying the electric conductivity mechanism. He also points out that to a considerable degree, electric conductivity is dependent on small impurities in rock samples and the degree of their oxidation, caused by the medium surrounding the sample during the experiment.

Such is the fundamental information on changes in the electric conductivity of rocks under different thermodynamic conditions, the object of our investigations.

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ELECTRIC PROPERTIES OF ROCKS
AT HIGH TEMPERATURES AND PRESSURES

Measurement Apparatus and Method

In determining the resistivity of rocks we tested several forms of appara-   /10
tus and systems with a MOM-4 or E6-3 thermaohmmeter. The measurements were
made by the dc current method at a high temperature and at atmospheric pressure
in a muffle furnace with a maximum working temperature up to 1,250°. Contact
with the sample was with two flat electrodes which were attached by a special
spring device. Temperature was measured with platinum-platinum-rhodium thermo-
couples. The samples were prepared in the form of disks 15 mm in diameter and
5 mm high.

In studying the effect of unilateral pressure at room temperature and with
heating to 250° we used "bombs" of the Adams type with external heating. A
sample 15 mm in diameter and up to 20 mm high was packed in a pyrophyllite or
plastic sleeve. Punches of instrument steel at the same time served as elec-
trodes. The magnitude of unilateral pressure attained 20,000 kg/cm^2; the com-
puted hydrostatic pressure was approximately 2,000 to 3,000 kg/cm^2.

Figure 1 is a diagram of a high-pressure apparatus used in measuring re-
sistivity at a temperature ~ 600° and a pressure ~ 30,000 kg/cm^2. The cylinder,
punches and punch supports were fabricated from thermally processed high-speed
R-18 steel; the cylinder and supports supporting elements were made from 40X   /11
alloyed steel. The investigated sample was placed in a pyrophyllite sleeve
which fitted tight in the cylinder. Gaskets of thin copper foil were used for
improving the electric contact between the sample and the punches. The electric
insulation of the measuring circuit was insured by a set of mica gaskets. The
thermocouple measuring leads passed through porcelain and quartz tubes. During
the course of the experiment the temperature of the lateral surface of the
apparatus was monitored. The force on the punches was created by a 220-ton
UVD-1 press.

In order to study the effect of temperature on resistivity of some types
of igneous rocks at normal pressure the samples were heated to 1,200°. Resis-
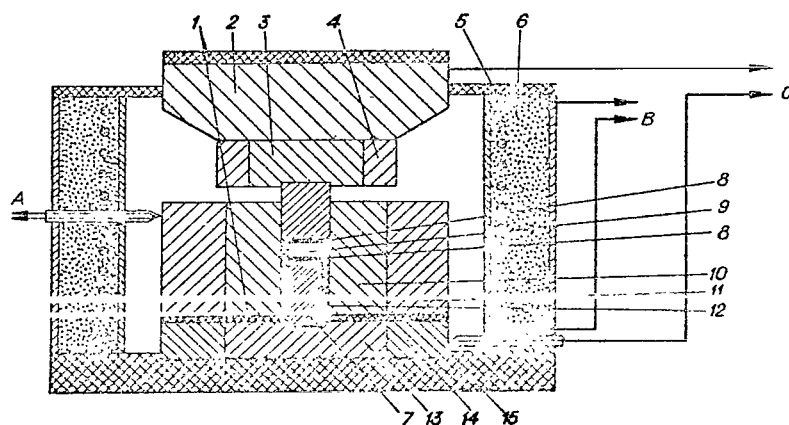
Figure 1. High-pressure apparatus for determining rock resistivity.

1 - insulation; 2 - piston; 3 - upper support; 4 - supporting unit for supports; 5 - furnace winding; 6 - furnace; 7 - punches; 8 - copper foil; 9 - investigated sample; 10 - working cylinder; 11 - support for cylinder 10; 12 - pyrophyllite sleeve; 13 - lower support; 14 - supporting unit for support; 15 - textolite base; A - thermocouple; B - current supply for furnace; C - electric leads.

tivity was measured during heating and cooling and also during repeated heating.

In studying the effect of primarily unilateral pressure up to 20,000 kg/cm^2 /12 (the hydrostatic pressure in these experiments was approximately 2,000 to 3,000 kg/cm^2) most of the experiments were performed at room temperature. In individual cases the rock samples were heated, but by not more than 250°.

During the subsequent study of rock resistivity the experiments were conducted under the simultaneous influence of temperature and pressure. The hydrostatic pressure attained 30,000 kg/cm^2 at a temperature ∼ 500°.

Experiments for the study of rock electric conductivity were performed in three regimes: (1) at normal atmospheric pressure, but at high temperatures; (2) at room temperature and increased (unilateral) pressures; and (3) at high temperatures and pressures.

Electric Conductivity of Rocks at High Temperatures

At a temperature up to 1,200° we studied the resistivity of granite, olivinite, dunite, and pyroxenite. The measurement results are given in Figure 2 and Table 1.

The resistivity values for the studied dunite samples are inconstant prior to heating and vary in the range $1 \cdot 10^9$ - $5 \cdot 10^9$ ohm·m. At the maximum heating temperature almost all the values coincided, attaining $4.2 \cdot 10^2$ - $4.6 \cdot 10^2$ ohm·m. The resistivity values for dunite samples after cooling, like for other rocks, were lower than the initial values: $8.2 \cdot 10^8$-$1.7 \cdot 10^9$ ohm·m.

Individual pyroxenite samples differed in resistivity prior to heating ($1 \cdot 10^9$ - $2 \cdot 10^{10}$ ohm·m), at maximum temperature ($5.0 \cdot 10$ - $2.7 \cdot 10^2$ ohm·m), and after cooling ($2.1 \cdot 10^7$ - $3.5 \cdot 10^8$ ohm·m). A similar picture is also noted for the other investigated rocks. The curves in Figure 2 illustrate the general nature of the changes in resistivity for the investigated rocks as a function of temperature variations. (The figure scale corresponds to the functional relationship $\ln \rho = f(1/T)$, where T is absolute temperature).

Analysis of these curves makes it possible to detect some peculiarities in the change of resistivity of rocks at high temperature, characteristic of all the investigated rocks.

First there is a regular decrease in resistivity in the process of heating /13 to 1,200°, attaining 6 to 8 orders of magnitude in comparison with the initial level. The resistivities at 1,200° are extremely close for different rocks.
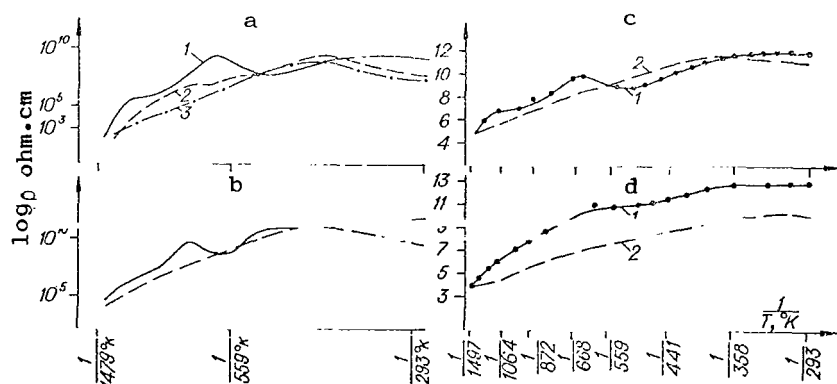

Figure 2. Dependence of resistivity of granite (a), pyroxenite (b), dunite (c), and olivinite (d) on temperature.

1 - with initial heating; 2 - with cooling; 3 - with repeated heating.

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Despite the decrease in resistivity with a temperature increase, the resistivity curves have local peaks, "humps," which are observed to a temperature of 500°. At 100 to 110° the heating curves frequently show a resistivity peak which is clearly repeated on the cooling curves. The resistivity peak on the heating curve, observed in the range 350 to 500°, disappears during cooling. At a higher temperature resistivity decreases monotonically; the measured resistivities fall on a straight line.

In most cases the resistivity curve for the first heating does not coincide with the cooling curves and the repeated heating curves. The ordinates of the latter are less than the ordinates of the first curve and the curves themselves are considerably straightened.

TABLE 1. RESISTIVITY OF ROCKS AS A FUNCTION OF TEMPERATURE

Rock	Resistivity, ohm·m		
	Before heating	Maximum heating (1,200°)	After cooling
Dunite	$5.3 \cdot 10^9$	$4.2 \cdot 10^2$	$1.7 \cdot 10^9$
"	$5.0 \cdot 10^9$	$4.4 \cdot 10^2$	$8.2 \cdot 10^8$
"	$1.1 \cdot 10^9$	$4.6 \cdot 10^2$	$1.3 \cdot 10^9$
Olivinite	$8.5 \cdot 10^{10}$	$2.8 \cdot 10$	$1.1 \cdot 10^8$
"	$4.0 \cdot 10^9$	$8.5 \cdot 10$	$2.8 \cdot 10^7$
Granite	$5.3 \cdot 10^7$	$4.4 \cdot 10$	$1.4 \cdot 10^7$
"	$1.3 \cdot 10^9$	$8.8 \cdot 10$	$4.6 \cdot 10^7$
Pyroxenite	$1.3 \cdot 10^9$	$5.0 \cdot 10$	$5.5 \cdot 10^8$
"	$2.1 \cdot 10^{10}$	$2.7 \cdot 10^2$	$1.9 \cdot 10^8$
"	$3.4 \cdot 10^9$	$1.1 \cdot 10^2$	$2.1 \cdot 10^7$

Electric Conductivity of Rocks Under Unilateral Pressure

Under unilateral pressure up to 20,000 kg/cm^2 and hydrostatic pressure not exceeding approximately 2,000 to 3,000 kg/cm^2 at room temperature and sometimes with heating to 250° we studied the resistivity of olivinite (Monchegorsk), marble (without site identification), serpentinite (Urals, Eastern Sayan), dunite (Urals), basalt, pyroxenite, and peridotite (Urals). Figure 3 shows curves of the dependence of resistivity on pressure for rocks of different

composition. Under the influence of pressure all the investigated rocks exhibit a decrease in resistivity, attaining some minimum value, and then again increasing. The nature of the examined curves is different for each rock variety. The resistivity decrease varies in the range from fractions of one to two orders of magnitude. The greatest resistivity change was observed for marble, serpentinite and basalt (curves 1-3); the minimum was observed for peridotite and pyroxenite (curves 5, 6). The minimum on the resistivity curve for each rock type corresponds to a different pressure on the sample. The minimum value of the latter is 700 kg/cm^2 for serpentinite and the maximum value is 8,500 kg/cm^2 for dunite. The interval of the minimum values on the resistivity curves is expressed differently for different rocks. For basalt, serpentinite and marble the minima have the form of a sharp peak. After transition through the minimum their resistivities increase sharply, attaining values exceeding the initial levels. The minimum resistivity value for pyroxenite, which remains constant, at pressures of 1,400 to 12,000 kg/cm^2 corresponds to an almost straight line. In the pressure range 12,000 to 20,000 kg/cm^2 it increases stably with a uniform resistivity increment exceeding 1 to $1.5 \cdot 10^7$ ohm·m per 1,000 kg/cm^2. For the other rocks which we studied the resistivity change is characterized by intermediate resistivity values for basalt, serpentinite and pyroxenite.

The following conclusion can be drawn on the basis of the above:

(a) with a pressure increase the resistivity of rocks initially decreases;

(b) the pressure at which the minimum resistivity is attained varies from 700 to 8,500 kg/cm^2 for different rock types;

(c) the minima on the curves of the dependence of resistivity on pressure are differently expressed: in some cases in the form of sharp peaks, and in others in the form of more indistinct peaks with a horizontal segment in a considerable pressure range;

(d) with a further pressure increase, resistivity increases, to different degrees for different rocks (see Figure 3).

In connection with the data given above on the changes of rock resistivity with a change in pressure it is of interest to consider some information on the behavior of volumetric weight and specific gravity, porosity and structure of these same rocks under these same conditions. However, due to the lack of

TABLE 2. SPECIFIC GRAVITY, VOLUMETRIC WEIGHT AND COEFFICIENT OF TOTAL ROCK POROSITY.

| Rock	Specific gravity, g/cm³	Density g/cm³	Porosity %	Specific gravity, g/cm³	Density g/cm³	Porosity, %	Pressure, kbar	Temperature °C
	Under normal conditions			After exposure to pressure and temperature				
Pyroxenite	3,34	3,27	2,09	3,38	3.03	10,3	13,8	20
				3,33	2,90	12,7	14,0	20
				3,33	2,98	10,5	25,5	20
Dunite	2,90	2,84	2,07	2,90	2,46	15,1	11,2	20
				2,88	2,60	9,7	11,8	20
				2,90	2,34	19,3	16,6	20
Peridotite	2,81	2,76	1,77	2,82	2,36	16,3	12,4	20
				2,81	2,48	11,7	13,8	20
Olivinite	3,44	3,36	1,70	3,42	3,03	14,2	13,1	20
					3,00	11,2	19,1	20
Serpentinite	2,73	2,65	2,93		2,77	5,1	4,1	20
					2,38	9,4	16,6	20
					2,58	7,8	17,2	20
				2,73	2,44	10,6	22,6	250
Eclogite	3,31	3,30	0,30	3,29	2,96	10,0	21,0	20

Commas represent decimal points

ohm·m
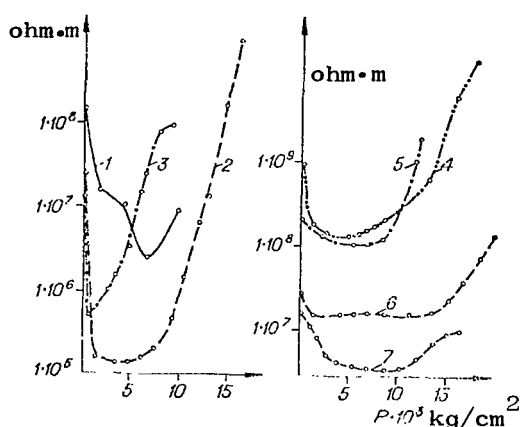
Figure 3. Dependence of rock resistivity on unilateral pressure.

1 - basalt; 2 - marble; 3 - serpentinite; 4 - olivinite;
5 - peridotite; 6 - pyroxenite;
7 - dunite.

special apparatus and measurement methods for determining these parameters, they were determined during the course of the experiment, under normal conditions and after pressure was removed.

Specific gravity was determined by the hydrostatic weighing method in pure alcohol in micropycnometers and volumetric weight was determined by the weighing of paraffined samples in distilled water (Table 2). Only the specific gravity of rocks did not change under the influence of the pressure to which the rock was subjected (the observed specific gravity variations were within the limits of measurement error).

However, the volumetric weight of all the rocks decreased considerably, obviously caused by an increase in their total porosity coefficient. The decrease in volumetric weight and increase in rock porosity evidently occurred because unilateral pressure predominated in the experiments and the rocks cracked when it was applied. For the time being it has been impossible to establish a more precise relationship between the degree of pressure, on the one hand, and the porosity coefficient or volumetric weight on the other. For example, for olivinite at a pressure of 19,000 kg/cm^2 the total porosity coefficient was 11.2% and at 13,000 kg/cm^2 it was 14.2%, although it would seem that the reverse should be true. Similar examples can also be given for other rocks. This matter obviously requires special investigation.

Electric Conductivity of Rocks at High Temperatures and Pressures

The last stage in the described experiments was a series of experiments at high temperatures and pressures. As already mentioned, in investigating the influence of unilateral pressure in Adams "bombs" some of the samples were heated to 250°. However, in the high-pressure apparatus whose design is shown in Figure 1, rock resistivity was determined with their heating above 500° and at a hydrostatic pressure up to 31,000 kg/cm^2. Under these conditions we studied the behavior of olivinite (Monchegorsk), serpentinite (Urals), and

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eclogite (Northern Kazakhstan). The olivinite and eclogite samples were
heated to 600°, after which they were subjected to a pressure up to 23,000
kg/cm^2. The heating temperature was maintained constant during the experiment.
The behavior of serpentinite was studied at 440° and a pressure of 31,000 kg/cm^2.
Figure 4 shows that all rocks were characterized by a resistivity decrease. How-
ever, the degree of this decrease and the nature of the change were different
for different rocks. For example, at atmospheric pressure and a temperature of
600°C the resistivity of olivinite was $5.6 \cdot 10^6$ ohm·m (in Figure 4 this point
corresponds to the beginning of curve 1). The pressure increase at this same
temperature caused a smooth decrease in olivinite resistivity; at a pressure
of 10,000 kg/cm^2 it attained $6.4 \cdot 10^5$ ohm·m, and at a pressure of 23,000 kg/cm^2
it was $3.3 \cdot 10^5$ ohm·m.

A somewhat different picture was observed for eclogite, whose resistivity
at a temperature of 600° and atmospheric pressure was $5.3 \cdot 10^5$ ohm·m. With a
pressure increase to 4,500 kg/cm^2 it decreased sharply (to $2.3 \cdot 10^5$ ohm·m). With
a further increase in pressure to 17,000 kg/cm^2 the resistivity of eclogite had
a tendency to a slow increase (see Figure 4).
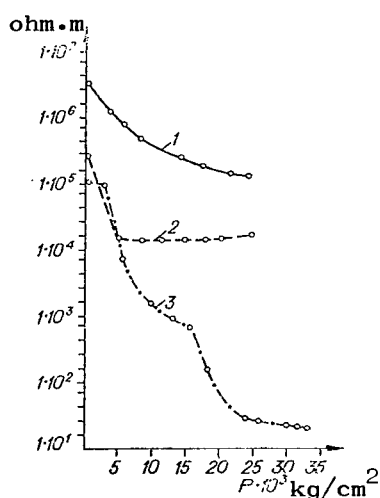
Figure 4. Dependence of resistivity of olivinite (1), eclogite (2) and serpentinite (3) on pressure at high temperatures.

In the experiments with serpentinite the heating temperature was somewhat lower (440°), but the effect of a change in its resistivity was mani- fested considerably more clearly: the range of de- crease attained four orders of magnitude. At the same time (see Figure 4), for serpentinite with a pressure increase from 2,500 to 31,000 kg/cm^2, an extremely significant (by four orders of magnitude) two-step decrease in resistivity was characteristic. A very insignificant (from $1 \cdot 10^5$ to $9.5 \cdot 10^4$ ohm·m) decrease in resistivity at a pressure up to 2,500 kg/cm^2 is then replaced by a sharp (by a factor of 10^2) crease in the pressure range 2,500 to 9,000 kg/cm^2. At pressures of 9,000 to 15,000 kg/cm^2. resistivity again decreases slowly, by only 0.3 to $0.5 \cdot 10^3$ ohm·m per 1,000 kg/cm^2, after which in the pressure range 15,000 to 21,000 kg/cm^2 there is

again a sharp resistivity drop-off from $8.5 \cdot 10^2$ to $5.5 \cdot 10$ ohm·m. A further pressure increase (to 27,000 kg/cm^2) results in a resistivity decrease by an extremely small value. At the maximum pressure (31,000 kg/cm^2), attained in experiments with serpentinite, its resistivity was 42 ohm·m, that is it decreased by four orders of magnitude from the initial level at room temperature and atmospheric pressure.

These investigations revealed interesting peculiarities in the change in rock resistivity as a function of temperature and pressure. In particular, in all the investigated rocks it decreases with a temperature increase and at normal pressure. It is important to note that for rocks of different composition differing in resistivity at room temperature, the resistivities become close in value at a temperature $\sim 1,000°$ (see Table 1).

Under the influence of predominantly unilateral pressure at 20° the nature of changes in rock resistivity is considerably more complex: with a pressure increase it decreases to definite levels and then increases, sometimes exceeding the initial level.

Under high temperatures there is a constant decrease in resistivity with an increase in hydrostatic pressure. These results must be taken into account when interpreting variations of the geomagnetic field, when studying the electric conductivity of rocks in the deep layers of the earth.

Rock Electric Conductivity Mechanism

Igneous rocks constitute a hard mineral aggregate whose electric and thermal properties are determined by its composition, the nature of the chemical bond and texture. Rock texture is of three types: polycrystalline, vitreous (amorphous) and hypocrystalline. The latter contains, in different quantitative relationships, both the crystal structure elements and the uncrystallized vitreous residue.

The mineral composition of igneous rocks is represented for the most part /20 by oxides; more than 74% is accounted for by SiO_2 and Al_2O_3 (Zavaritskiy, 1955), each of which in pure form is characterized by high resistivity. For example, according to data published by W. D. Kingery (1963), the resistivity of crystalline sapphire (99.9% Al_2O_3) at 20° is more than 10^{12} ohm·m and at 1,000° is 10^6 ohm·m, whereas the resistivity of quartz glass (99.8% SiO_2) at 20° is

above 10^{12} ohm·m and at 1,000° is 10^4 ohm·m.

In our experiments the samples of dunite, olivinite and other rocks at 20° also had a high resistivity (10^{10}-10^8 ohm·m). However, with heating of the sample to 1,000° the resistivity decreased to 10-10^2 ohm·m. The relatively low rock resistivities at high temperatures are probably caused by different impurities in the crystalline grains and intercrystalline layer. The impurities can favor an increase in the concentration of charge carriers and extrinsic conductivity appears in rocks which is not present in pure minerals (for example, in the mentioned sapphire). It should be noted that for the time being it is impossible to say anything definite concerning the relative role of conductivity of crystalline grains and the intercrystalline layer. Accordingly, it is better not to distinguish between them and instead use for igneous rock a hypothetical model of a contaminated semiconductor with a broad forbidden zone. However, in formulating such a model it is extremely important that a large number of impurities is assumed, leading to a very low resistivity of igneous rocks, which in actuality is not observed. This difficulty disappears if it is assumed that the rock is a semiconductor having mixed conductivity and this conductivity is highly compensated, that is, the impurities of the donor and acceptor types have almost identical concentrations.

This model makes it possible to explain the experimental data. The curves of the dependence of resistivity on temperature during the initial heating of the rock samples are complicated by a number of local maxima, "humps." Each "hump" can be attributed to the presence of impurities with a definite activation energy and a system of "humps" can be attributed to the presence of several types of impurities with different activation energies which serve as additional /21 charge carriers for different temperatures, determined by the activation energy.

At high temperatures there are irreversible processes with a restructuring of the extrinsic atoms. As a result, the cooling curve is smoothed, assuming the shape of an almost straight line (see Figure 2). This can be described by the empirical formula

$$\rho = CT^{\beta} \exp{-\frac{E}{KT}}$$

assuming E is constant. In the formula C and β are constants; K is the Boltzmann constant and the parameter E can be regarded as the activation energy.

If the dependence of ρ on T is experimentally determined, this formula can be used in estimating the activation energy of impurities in the investigated rock. The determined β and E values for some rocks are given in Table 3. The activation energy for all the examined rocks is less than 1.0 eV. However, it is known that for pure minerals it is considerably greater (for example, the Al_2O_3 activation energy is more·than 5 eV). Thus, the order of the determined E values indicates extrinsic conductivity in the igneous rocks.

TABLE 3. ACTIVATION ENERGY OF SOME IGNEOUS ROCKS
AT DIFFERENT TEMPERATURES

Rock	Activation energy E, eV	Parameter β	Temperature range °C	Remarks
Dunite	0.91	0.32	360-1140	Heating
"	0.97	0.05	1170-140	Cooling
"	0.77	O	1060-500	"
Olivinite	0.47	O	920-100	"
"	0.69	0.75	790-160	"
Pyroxenite	0.85	0.02	1060-160	"

This model, although it makes it possible to explain the results of experiments for determining the dependence of resistivity of igneous rocks on temperature, nevertheless is not reliable for a quantitative description of these phenomena. A large quantity of similar data and special investigations in this field are obviously necessary.

CHAPTER II

THERMAL PROPERTIES OF ROCKS AT ROOM TEMPERATURE

Measurement Apparatus and Method

Nonstationary and stationary methods were used in measuring the thermal /22
coefficients as a function of the size of the rock sample. In the first case
the thermal properties were measured on large samples (with a minimum distance
of 100 mm between surfaces) by the pulse probe method developed by G. N.
Starikova and A. P. Shushpanov and described in detail in Geotermicheskiye
Issledovaniya [Geothermal Investigations] (Nauka, 1964). We introduced only a
few modifications into this method. Figure 5 is a diagram of the apparatus.
The probe method was also used in measuring cores 40-70 mm in diameter, but the
method was somewhat modified. In this variant the heater-spring was replaced by
a constantan wire 0.2 mm thick which was placed in a through opening in the sample
(the diameter of the heater opening was 4 mm) and was held tight by fluoroplastic
wedges. The distance r between the heater and the thermocouple was 5-10 mm. The
thermocouple was placed in an aperture 4 mm in diameter parallel to the heater.
The thermojunction was at the level of the middle of the heater. Its good con-
tact with the rock was ensured by means of a device in the form of a wedge, as
in the case of the heater. The current source was a battery of dry cells with a
voltage of 8-10 V. The measuring unit remained the same as for large samples.
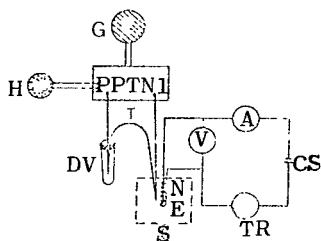
Figure 5. Diagram of apparatus
for determining thermophysical
constants by the probe method.
G- galvanometer; PPTN1- potentio-
meter; NE- normal element; DV-
Dewar vessel; T- thermocouple;
A- ammeter; V- voltmeter; TR-
time relay; CS- current source;
H- heater; S- sample.

The results of measurements of heat con- /23
ductivity λ made on large samples of white
marble and pyrophyllite, and then on small
samples prepared from them, agree well with
one another (Table 4).

The accuracy in measurements by the
probe method is 10%.

In determining heat conductivity on
homogeneous samples of very small size
(h = 6 mm, d = 12 mm) we used an A-25 in-
strument constructed by the Institute of
Metrology (Committee on Standards, Measures
and Measuring Instruments); its operation is based on the laws of a stationary

heat regime.

The heat flux q_x through the investigated sample can be written as follows:

$$q_x = \lambda_x \Delta t_x \cdot \frac{S_x}{h},$$

where

λ_x is the heat conductivity coefficient;

Δt_x is the temperature drop between the sample surfaces (the distance between them is h);

S_x is the area of the sample through which the flux passes.

TABLE 4.   VALUES OF THE HEAT CONDUCTIVITY COEFFICIENT
FOR WHITE MARBLE AND PYROPHYLLITE
DETERMINED ON LARGE AND SMALL SAMPLES

Rock	Heat conductivity of small sample (d = 40 mm), W/m·degree	Heat conductivity of large sample (d = 100 mm), W/m·degree
White marble	2.18	2.34
Pyrophyllite	3.10	3.27

The operating principle of the instrument is based on measurement of the   /24
temperature drop Δt_x with a constant heat flux. The instrument was calibrated
using a series of samples of identical size (h = 6 mm, d = 12 mm) with a known
heat conductivity coefficient (fused quartz, Plexiglas, pyrophyllite) using the dependence $\Delta t_x = f(\lambda_x)$ with q_x const. A calibration curve was constructed using the
results of these measurements. The measurement error did not exceed 8%.

Experimental Results

During the first stage in studying the thermal properties of rocks we
made a series of experiments at room temperature and at atmospheric pressure.
This gave us some idea concerning the characteristic heat conductivity values
for rocks of different composition and its relationship to other physical parameters and served as a point of departure for further similar but more complex
experiments at high temperatures. The laboratory investigation which we had
begun (Moiseyenko, Sokolova, 1963, 1967) was made using core samples from bore-

holes, ore lumps or natural exposures. We studied more than 500 samples of igneous and sedimentary rocks from Eastern Kazakhstan, the Eastern Sayan, Yuzhno-Minusinskaya depression, West Siberian Lowland, and the Kamchatkan Peninsula (Tables 5, 7).

Figure 6 shows the dependence of the heat conductivity coefficient on composition for intrusive rocks from Eastern Kazakhstan and the Eastern Sayan. The rocks are arranged along the x-axis in this figure in accordance with the degree in increase of their basicity; heat conductivity coefficient values λ are plotted along the y-axis. Each rock variety is characterized by a definite range of λ values, which for most of the studied rocks vary in a small range (the broadest range is for granodiorites, from 1.64 to 2.48 W/m·degree). Rocks of identical composition from the Eastern Kazakhstan and Eastern Sayan regions are characterized by close λ values. Its highest values (2.11-2.83 W/m·degree) are observed for alaskitic granites. The heat conductivity of leucocratic granites is lower and varies in narrower limits. The minimum values are observed for bifeldspathic granites. The λ results agree with data from other researchers for rocks of similar composition.

It also follows from Figure 6 that with an increase in basicity of the /25 rocks there is a decrease in their heat conductivity. The maximum λ value for granites is 2.83 W/m·degree, whereas for diorites it is 2.00 W/m·degree. This regularity does not extend to bifeldspathic granites, characterized by a lesser λ value in comparison with the more basic rocks. In porphyrites the λ value varies from 1.76 to 2.44 W/m·degree.

Tuffs of acidic and mixed composition from the Belousovskaya area (borehole 801) in Eastern Kazakhstan are characterized by variations in heat conductivity in the range 1.38-2.72 W/m·degree. The thermal conductivity K of these rocks varies from 1.02 to 4.23 m^2/hour and the heat capacity C from 0.19 to 0.38 Cal/kg·degree (see Table 5).

With an increase in fissuring and weathering (which leads to an increase in porosity) the heat conductivity of rocks is decreased (see Table 5).

The same regularity is observed in samples of hardened lava from the Karymskiy Volcano on Kamchatka, constituting a hard, very porous material.

TABLE 5. THERMAL PROPERTIES OF IGNEOUS ROCKS
IN EASTERN KAZAKHSTAN AND THE EASTERN SAYAN

Rock	Sampling site (granite complex)*	Heat conductivity λ W/m·degree	Thermal conductivity $K \times 10^{-3}$, m^2/hour	Heat capacity C, Cal/kg	Density γ, g/cm^3	Interval of removing core from borehole, m
Alaskitic granite, coarse grained	Kandygatay	2.75	2.5	0.26	2.60	-
Same	"	2.71	2.1	0.33	2.65	-
"	"	2.70	2.5	0.33	2.60	-
Alaskitic granite (from dike)	Koytas	2.61	1.8	0.32	2.63	-
"	Shindinskiy	2.83	4.2	0.21	2.60	-
"	Kandygatay	2.11	3.4	0.33	2.62	-
Alaskitic granite coarse-grained, yellow, highly weathered (crumbles)	"	1.65	1.9	0.21	2.62	-
Alaskitic granite fine-grained	"	2.69	-	-	-	-
Alaskitic granite, fine-grained, (from dike)	Kalba	2.08	1.4	0.30	2.61	-
Leucocratic granite intermediate grain (from dike)	Koytas	1.62	1.7	0.24	2.61	-
Leucocratic granite coarse-grained	Kandygatay	2.26 2.30	- 3.1	- 0.26	- 2.60	- -
Leucocratic granite fine-grained	Koytas	2.16	2.1	-	-	-

Table 5. Continued

Rock	Sampling site (granite complex)*	Heat conductivity λ W/m·degree	Thermal conductivity $K \times 10^{-3}$, m^2/hour	Heat capacity C, Cal/kg	Density γ g/cm^3	Interval of removing core from borehole, m
Leucocratic granite, (from dike)	Koytas	2.27	-	-	-	-
Leucocratic granite, coarse grained, highly weathered	Kandygatay	2.12	-	-	-	-
Leucocratic granite, coarse-grained	Shindinskiy	2.38	2.5	0.26	2.63	-
Leucocratic granite, intermediate grain	"	2.40	2.1	0.28	2.63	-
Leucocratic granite coarse-grained	"	2.36	3.1	0.23	2.63	-
Leucocratic granite	Kandygatay	2.29	-	-	2.62	-
Biotite granite coarse-grained	Ulen'-Tuim	2.05	2.4	-	-	-
Biotite-horneblende granite intermediate grain	Koytas	2.29	-	-	2.59	-
Biotite-horneblende silicified granite	Shindinskiy	2.56	2.4	-	-	-
Pyroxene granite, fine-grained	Ulen'-Tuim	2.00	2.3	-	-	-
Modified granite, coarse-grained	Shindinskiy	2.39	2.2	-	-	-
Alaskitic granite-porphyry	Ulen'-Tium	2.26	1.8	0.33	2.64	-
Porphyraceous granite	Kandygatay	2.17	-	-	-	-
Porphyraceous granite (from dike)	Kalba	2.27	3.6	0.25	2.56	-

Table 5. Continued

Rock	Sampling site (granite complex)*	Heat conductivity λ W/m·degree	Thermal conductivity $K \times 10^{-3}$, m^2/hour	Heat capacity C, Cal/kg	Density γ, g/cm^3	Interval of removing core from borehold, m
Bifeldspathic granite	Kalba	1.49	-	-	-	-
Same	"	1.65	-	-	2.66	-
"	Koytas	1.73	1.5	0.28	2.60	-
Bifeldspathic granite	Kandygatay	1.68	1.2	0.33	2.62	-
Leucocratic plagiogranite fine-grained (from dike)	Shindinskiy	2.03	3.0	-	-	-
Granodiorite intermediate grain	Koytas	2.48	2.0	-	-	-
Granodiorite	Koytas	2.08	2.2	0.30	2.62	-
"	"	1.64	1.4	0.26	2.65	-
"	"	1.87	-	-	-	-
"	Kalba	2.09	1.7	-	-	-
Leucocratic granodiorite intermediate grain	"	1.95	1.5	0.30	2.66	-
Porphyraceous granodiorite	"	2.05	1.5	-	-	-
same	Kandygatay	1.77	-	-	-	-
Granodiorite intermediate grain	Tigertyshskiy	2.10	2.0	0.28	2.76	-
same	"	1.91	2.5	0.204	2.75	-
"	Shindinskiy	2.29	1.9	-	-	-
"	"	2.06	1.9	0.25	2.73	-
"	"	2.27	1.9	0.3	2.75	-
"	"	3.31	2.7	-	-	-
Granodiorite coarse-grained	Kalba	1.96	1.9	-	-	-
Porphyraceous granodiorite fine-grained	Shindinskiy	1.91	2.5	0.22	2.69	-

Table 5. Continued

Rock	Sampling site (granite complex)*	Heat conductivity λ W/m·degree	Thermal conductivity $K \times 10^{-3}$, m²/hour	Heat capacity C, Cal/kg	Density γ, g/cm³	Interval of removing core from borehole, m
Granodiorite fine-grained	Shindinskiy	1.95	1.1	0.24	2.75	-
Diorite, intermediate grain	"	1.91	2.6	-	-	-
Horneblende diorite	"	1.81	2.1	-	-	-
Diorite-porphyry	Ulen'-Tuim	2.00	1.9	0.27	2.87	-
"	"	1.80	1.8	0.26	2.88	-
"	Kandygatay	1.91	1.2	0.38	2.69	-
Gabbro	Ulen'-Tuim	2.27	1.9	-	-	-
Diabase fine-grained	Tigertyshskiy	1.97	2.1	-	-	-
Diabase, silicified	Shindinskiy	2.35	2.9	-	-	-
Contact-modified diabase with ore impregnation	"	2.00	2.5	-	-	-
Dike porphyrite	Uybatskiy	1.96	1.8	-	-	-
Quartz porphyry	Kandygatay	1.76	-	-	-	-
Quartz porphyry, highly recrystallized	"	2.44	-	-	-	-
Quartz porphyry	"	1.98	1.8	0.28	2.64	-
same	"	1.77	-	-	-	-
Andesitic horneblende porphyry, intermediate composition	Koytas	2.26	-	-	-	-
Marmorized limestone	Shindinskiy	2.77	2.6	-	-	-

Table 5. Continued

Rock	Sampling site (granite complex)*	Heat conductivity λ W/m·degree	Thermal conductivity $K \times 10^{-3}$, m^2/hour	Heat capacity C, Cal/kg	Density γ, g/cm^3	Interval of removing core from borehold, m
Horneblende tuff	Kandygatay	2.37	-	-	-	-
Porphyraceous tuff	Kalba	2.11	4.2	0.19	2.69	-
Tuff, mixed composition	Eastern Kazakhstan	2.18	-	-	-	185-191
same	Belousovskaya area	1.60	-	-	-	241-243
same	Borehole 801	1.60	-	-	-	289-290
same	"	2.59	-	-	-	302-304
Tuff, mixed composition at boundary with tuffs of acidic composition	"	2.00	-	-	-	304-307
Tuff of acidic composition	"	1.88	-	-	-	321-323
same	"	1.72	-	-	-	326-328
Tuff of mixed composition	"	2.34	-	-	-	396-399
same	"	2.26	-	-	-	603
"	"	2.13	-	-	-	700-702
"	"	2.39	-	-	-	-

*Kandygatayskiy, Koytasskiy, Kalbinskiy granite complexes are in Eastern Kazakhstan; Shindinskiy, Ulen'-Tuimskiy, Tigertyshskiy, Uybatskiy granite complexes are in the Eastern Sayan.
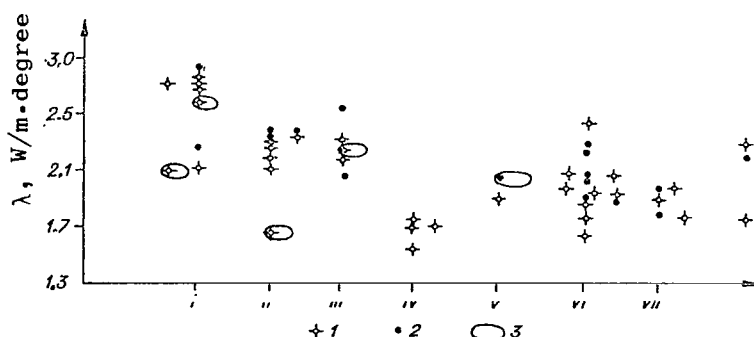
Figure 6. Heat conductivity coefficient for igneous rocks.

I - alaskitic granites; II - leucocratic granites; III - biotite-horneblende granites; IV - bifeldspathic granites; V - plagiogranites; VI - granodiorites; VII - diorites and dioritic porphyries; 1 - rock samples collected in Kazakhstan; 2 - rock samples collected in the Sayans; 3 - samples from dikes.

As might be expected, we obtained low values for their heat and thermal con- /31 ductivity coefficients, as can be seen from Table 6 which gives both the values of these coefficients and the densities of lavas. Coefficients K and λ increase with an increase in density (and consequently with a decrease in porosity).

TABLE 6. HEAT CONDUCTIVITY, THERMAL CONDUCTIVITY
AND HEAT CAPACITY COEFFICIENTS AND DENSITY OF LAVAS
FOR KARYMSKIY VOLCANO (KAMCHATKA)

Number of sample	Heat conductivity γ, W/m·degree	Thermal conductivity K, x 10^{-3}, m^2/hour	heat capacity C, Cal/kg·degree	Density γ, g/cm^3
527	0.25	0.84	0.27	0.95
407	0.33	0.94	0.28	1.10
517	0.48	0.85	0.33	1.47
521	0.65	1.09	0.30	1.76
503	0.73	1.49	0.16	2.64

The data given in Table 5 applied to the igneous rocks of the Eastern Sayan, Eastern Kazakhstan and Kamchatka. Table 7 gives the results of measurements for sedimentary rocks of the Minusinskiy downwarp, West Siberian platform and the Kronotskiy Peninsula on Kamchatka, taken from deep boreholes.

The Devonian rocks of the Minusinskiy downwarp are stratified with calcite veinlets, primarily gray and reddish aleurolites (siltstones), argillites, sandstones, and limestones. The heat conductivity of aleurolites varies in a

TABLE 7. HEAT CONDUCTIVITY COEFFICIENT FOR SEDIMENTARY ROCKS

Characteristics	Sampling site	Sampling interval, m	Heat conductivity λ, W/m·degree
Argillite, brownish-cinnamon, cemented	Yuzhno-Minusinskaya depression	805-811	2.68
Argillite, grayish-green, more or less stratified	Bystryanskaya area, borehole 9	826-831.8	2.13
Argillite, dark brownish-cinnamon, calcareous, more or less stratified	Same	931.9-937	1.80
Argillite, chocolate-cinnamon with inclusions of greenish aleurolite	"	961-971	2.89
Aleurolite, dark gray, more or less stratified, contains calcite veinlets	"	1248.9-1254	1.89
Argillite, black, more or less stratified, highly calcareous, contains calcite veinlets	"	1701-1710	2.68
Aleurolite, dark gray, with green hue	"	1852-1855	1.63
Aleurolite, gray, fine-grained dense, highly calcareous	"	1899-1900	2.01
Aleurolite, dark gray, noncalcareous, contains small calcite intercalations	"	1915-1917	1.80
Aleurolite, dark gray to black, undulatingly stratified	"	1945-1950	1.65
Aleurolite, calcareous	Bystryanskaya area, borehole 9	2051.6-2958	1.55
Aleurolite, brownish-cinnamon. more or less stratified, with gypsum inclusions	Same	2068-2073	2.26
Aleurolite, brownish-cinnamon,	Bystryanskaya area, borehole 9	2787-2794	1.80
Aleurolite, gray, highly calcareous	Same	2174.7-2182	1.67

Table 7. Continued

Characteristics	Sampling site	Sampling interval, m	Heat conductivity λ, W/m·degree
Sandstone, rosy-cinnamon, cemented, calcareous, contains calcite intercalations	Bystryanskaya area, borehole 9	1282-1290	2.13
Limestone, dark gray to black, with inclusions of crystalline calcite	"	1995-2000	2.09
Limestone, dark gray, dense, more or less stratified	"	2012-2018	2.30
Porphyrite, greenish-gray, strong	"	3159-3162	2.30
Aleurolite, reddish-cinnamon	Borehole 15	983.3-992.3	1.80
Aleurolite, gray	"	1904.7-1907	1.67
Aleurolite, dark gray	"	2183.8-2184.9	1.67
Aleurolite, gray, fine-grained	"	2211.5-2214.0	1.76
Sandstone	"	2163.8-2171.6	2.18
Aleurolite	Borehole 4	1240.9-1241.6	1.97
"	"	1235.7-1240.9	2.01
"	"	1699.8-1708.7	1.76
"	"	1880.4-1882.5	1.34
Sandstone	Altayskaya area, borehole 11	1023-1028.5	2.34
Sandstone, cinnamon-gray, fine-grained, very strong, low calcareousness	Same	1062.6-1072.7	2.05
Same	"	1062.6-1072.7	2.05
Sandstone, cinnamon-gray, very strong, calcareous	"	1675.6-1681.7	2.30
Sandstone, dark gray, calcareous, with cracks	Altayskaya area, borehole 11	1862.9-1870.0	2.05
Sandstone, light gray, very strong	Same	2157.8-2162.3	2.76
Same	"	2157.8-2162.3	2.76

Table 7. Continued

Characteristics	Sampling site	Sampling interval, m	Heat conductivity λ, W/m·degree
Limestone, dark gray, strong, with inclusions of white anhydrite˙	Altayskaya area, borehole 11	2162.2-2171.5	2.64
Same	"	2162.2-2171.5	2.72
Argillite, dark gray, slightly calcareous, strong	"	2112-2115.9	2.76
Argillite, gray, very strong, with calcite veinlets	"	2290.6-2292.6	3.01
Aleurolite, highly calcareous, with calcite veinlets	"	2294.7-2297.5	2.89
Argillite, dark gray, almost black, horizontally stratified, strong	"	2301.5-2305.2	2.47
Aleurolite, reddish-cinnamon, calcareous, horizontally stratified	"	2424.7-2428.7	2.26
Argillite, reddish-cinnamon, stratified	"	2462.2-2466.6	2.01
Aleurolite, gray	Sol'zavodskaya area, borehole 3	1818.3-1821.6	3.18
Aleurolite, gray, calcareous, fine-grained	"	1834.1-1837.5	1.67
Limestone, gray, intermediate grain	Same	1852.0-1855.1	2.22
Aleurolite, dark gray	"	1937.4-1941.3	2.00
Same	"	1941.3-1945.2	1.88
Porphyrite, cinnamon	"	2573.7-2582.0	1.72
Limestone, gray	"	1874.3-1878.9	2.64
Sandstone	Kamchatka, Kronotskiy region, Stolbovskaya structure, borehole GK6	225	1.63
Same	Same	270	1.47

29

Table 7. Continued

Characteristics	Sampling site	Sampling interval, m	Heat conductivity λ, W/m·degree
Sandstone	Kamchatka, Kronotskiy region, Stolbovskaya structure, borehole GK6	329	1.51
Same	Same	415	1.51
"	"	560	1.63
"	"	595	1.51
"	"	769	1.34
"	"	960	1.63
"	"	1060	1.84
"	"	1130	1.51
"	"	1179	1.80
"	Kamchatka, Kronotskiy region, Stolbovskaya structure, borehole GK5	410	1.58
"	Same	475	1.17
"	"	495	1.42
"	"	610	1.47
"	"	683	1.55
"	"	695	1.34
"	"	1110	1.72

/35

Note: Some samples of sandstone from Eastern Kamchatka (Kronotskiy region, Stolbovskaya structure) and samples of sedimentary rocks from the West Siberian Lowland were measured in a water-saturated state. The values of the heat conductivity coefficient for them are given in Table 8 (last column).

particularly wide range (1.3 to 3.2 W/m·degree). This is probably caused by an inhomogeneity in their composition and a large quantity of inclusions. The heat conductivity of argillites varies from 1.8 to 3.0 W/m·degree. For sandstones the variations in heat conductivity are insignificant. Its minimum value was 2.0 W/m·degree and its maximum value was 2.8 W/m·degree.

The investigated rocks of the Meso-Cenozoic sedimentary cover of the West Siberian Lowland are represented by sandstones, aleurolites and argillites. Their heat conductivity coefficient, as can be seen from Figure 7, varies from 0.58 to 1.80 W/m·degree. The broadest range of changes in this coefficient is /35 noted for sandstones, for which the entire above-mentioned range of values is characteristic. The heat conductivity of aleurolites varies from 1.31 to 1.53 W/m·degree. Most of the sedimentary rock samples for the West Siberian Lowland have a heat conductivity coefficient of 1.3 to 1.5 W/m·degree.
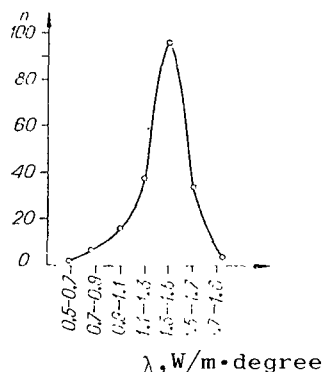

λ,W/m·degree

Figure 7. Variation curves for heat conductivity coefficient for sedimentary rocks from the West Siberian Lowland.

n – number of specimens.

The investigated samples from the Tertiary deposits of the eastern coast of the Kamchatkan Peninsula are coarse-, intermediate- and fine-grained sandstones. The grains are neither rounded nor crystallized and constitute fragments of different minerals. Their heat conductivity coefficient varies from 1.17 to 1.85 W/m·degree.

/36

The heat conductivity of sedimentary rocks is considerably affected by their porosity and the associated moisture content. A study of the dependence of heat conductivity on moisture content and porosity of rocks has been made by a number of researchers: W. Woodside and I. Messmer (Woodside, Messmer, 1961), T. Boldizsar (Boldizsar, 1964), B.A. Yakovlev and S.P. Vlasova (Sukharev, et al., 1966), E. Hurtig (Hurtig, 1966), and others.

T. Boldizsar (Boldizsar, 1964, 1965) found that for massive rocks with a porosity of about 5% or less the correction to heat conductivity for moisture content falls within the limits of measurement accuracy. Our experiments on igneous rocks (olivinite, eclogite, gabbro, diorite) led us to a similar conclusion.

However, if the investigated rock has considerable porosity, experimental data show that the influence of moisture content becomes significant and it must be taken into account.

W. Woodside and I. Messmer (Woodside, Messmer, 1961) investigated both unconsolidated sand and samples of dense quartzy sandstone. The porosity of the sandstones varies in a wide range from 3 to 59%. In order to clarify the degree of influence of the saturating fluid on effective heat conductivity the samples were saturated with fluids having different heat conductivities and a gas at a pressure of 1 atm. W. Woodside and I. Messmer also studied the dependence of heat conductivity on porosity. They found that heat conductivity increases with a decrease in porosity. With identical porosities the heat conductivity is higher for a sample which is saturated with a more heat conducting fluid.
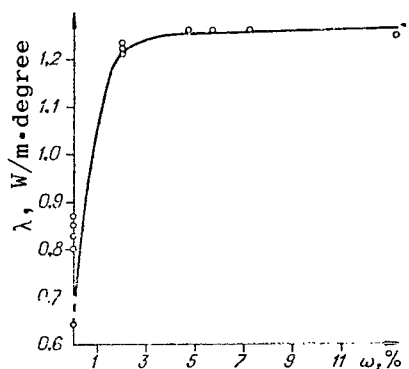
Figure 8. Dependence of heat conductivity coefficient of samples on moisture content ω .

We studied the dependence of heat conductivity /37 on porosity and moisture content for samples of sedimentary rocks collected in the Kronotskiy region on the Kamchatkan Peninsula and from the West Siberian Lowland (the lithological characteristics of these rocks were given above). Table 8 gives data on change in the heat conductivity coefficient for sedimentary rocks in these regions for different moisture contents and on their density and porosity.

The heat conductivity of most of the samples of sedimentary rocks was measured in two states: absolutely dry and with maximum water saturation. The heat conductivity of sandstones from the Kamchatkan peninsula was also measured on samples in an air-dry state. Figures 8 and 9 show the results of these heat conductivity measurements.

Figure 8 illustrates the nature of the dependence of heat conductivity of sandstones on moisture content (for boreholes in the Stolbovskaya structure on the Kamchatkan peninsula). A general pattern is observed: a marked increase in heat conductivity at the very beginning of moistening (approximately to 3% moisture content); then the rate of heat conductivity increase drops sharply.

TABLE 8. CHANGE IN HEAT CONDUCTIVITY
OF SEDIMENTARY ROCKS WITH MOISTENING

Sampling site (region, borehole)	Sampling interval, m	Rock	Density γ, g/cm³	Open porosity, %	Heat conductivity of an absolutely dry sample W/m·degree	Heat conductivity of sample in an air-dry state, W/m·degree	Heat conductivity of water-saturated sample, W/m·degree
Kronotskiy region, Eastern Kamchatka	248	sand stone	2,4I	I6,5	I,0I	I,24	I,27
	270	"	2,42	I5,3	0,87	I,04	I,12
Borehole GK6	329	"	2,29	I6,3	0,88	I,I0	I,I9
"	4I5	"	2,34	I6,2	0,95	I,I4	I,28
"	450	"	2,30	I6,6	–	I,27	I,29
"	50I	"	2,30	2I,4	0,77	I,03	I,25
"	590	"	2,34	I7,1	0,80	I,I7	I,26
"	595	"	2,33	I2,6	0,91	I,I9	I,26
"	890	"	2,36	I5,I	0,97	I,I6	I,34
Borehole GK5	263	"	2,20	I7,2	0,76	I,I4	I,25
"	330	"	2,22	I6,7	0,59	I,08	I,22
"	340	"	2,26	I9,4	0,77	I,12	I,2I
"	426	"	2,36	I4,8	0,72	I,I5	I,43
"	595	"	2,22	2I,4	0,72	I,5I	I,33
"	950	"	2,20	I7,8	0,72	I,25	I,27
West Siberian Lowland	2III	"	2,64	5,4	I,42	–	I,73
	2II4	"	2,06	22,7	0,98	–	I,35
Borehole 500	2I80	"	2,5I	4,9	I,76	–	I,43
Borehole 69	I940	"	2,29	I4,9	I,I6	–	I,45
Borehole 234	2I40	"	2,60	5,2	I,29	–	I,35
"	2I50	"	2,I8	25,4	I,I3	–	I,40
"	2545	"	2,24	I7,5	I,I3	–	I,49
Borehole 518	2060	"	2,55	5,9	I,48	–	I,50
Borehole 221	2II5	"	2,09	22,3	0,84	–	I,06
"	2I20	"	2,I9	I8,4	0,90	–	I,35
"	2I25	"	2,03	23,9	0,94	–	I,48
Borehole 547	2070	"	2,07	23,0	0,99	–	I,30
Borehole 74	2I00	"	2,65	2,6	I,04	–	I,27
Borehole 66	2070	"	I,98	25,6	0,83	–	I,40
Borehole 66	2I30	"	I,94	26,3	0,75	–	I,23
Borehole 504	2I20	"	2,05	23,I	0,89	–	I,44
"	2I25	"	I,98	27,I	0,94	–	I,29
"	2I35	"	2,03	24,2	0,8I	–	I,30

Commas represent decimal points.

Table 8. Continued

Sampling site (region, borehole)	Sampling interval, m	Rock	Density γ, g/cm³	Open porosity, %	Heat conductivity of an absolutely dry sample W/m·degree	Heat conductivity of sample in an air-dry state, W/m·degree	Heat conductivity of water-saturated sample, W/m·degree
Borehole 509	2060	sand	–	–	0,81	–	I,51
"	2I00	stone	2,00	24,I	0,94	–	I,19
Borehole 63	25⁴0	"	2,26	I6,3	I,08	–	I,40
Borehole 214	2080	"	2,03	23,5	0,8I	–	I,36
Borehole 236	2090	"	2,06	22,4	0,99	–	I,34
"	2III	"	I,99	27,I	0,94	–	I,29
"	–	"	2,64	3,6	I,27	–	I,51
Borehole 65	2044	"	2,00	24,2	0,89	–	I,39
"	2085	"	2,0I	25,2	0,9I	–	I,23
"	2090	"	I,99	24,2	I,03	–	I,30
"	–	"	2,I0	22,2	0,90	–	I,24
Borehole 549	I935	"	2,06	22,7	0,94	–	I,43
"	2095	"	2,00	24,5	0,94	–	I,40
Borehole 502	I980	"	I,96	27,I	0,9I	–	I,40
"	2I35	"	2,06	22,3	I,I2	–	I,43
Borehole 215	–	"	2,I0	20,7	I,I2	–	I,45
"	–	"	2,I8	I8,4	0,9I	–	I,36
Borehole 515	2620	"	2,44	I0,I	I,34	–	I,58
Borehole 72	–	"	2,45	I2,3	I,22	–	I,40
"	I885	"	2,I0	I9,7	0,89	–	I,I2
Borehole 116	2I50	"	I,98	25,2	0,94	–	I,I6
Borehole 88	2065	"	2,05	22,8	0,93	–	I,36
"	2II0	"	2,09	20,7	0,90	–	I,08
Borehole 67	2095	"	2,64	2,4	I,23	–	I,6I
Borehole 503	2I05	"	2,00	26,0	0,94	–	I,27
Borehole 81	2I20	"	2,05	23,4	0,83	–	I,I9
Borehole 69	I920	aleu-rolite	2,26	I7,4	I,I2	–	I,5I
"	I960	"	2,27	I6,0	I,30	–	I,5I
" "	2090	"	2,28	I5,6	0,95	–	I,48
"	2090	"	2,29	I6,0	I,08	–	I,39
"	2095	"	2,20	I5.5.	0,99		I,55
"	2I00	"	2,25	I7,4	I,I2	–	I,34
"	2I25	"	2,26	I7,8	I,II	–	I,54
"	2I30	"	2,33	I4,8	I,I6	–	I,45
Borehole 500	I875	"	2,24	I7,7	I,05	–	I,6I
"	I965	"	2,II	2I,8	I,06	–	I,39

Commas represent decimal points.

Table 8. Continued

Sampling site (region, borehole)	Sampling interval, m	Rock	Density γ, g/cm³	Open Porosity, %	Heat conductivity of an absolutely dry sample W/m·deg.	Heat conductivity of sample in an air-dry state, W/m·degree	Heat conductivity of water-saturated sample, W/m·degree
Borehole 234	2060	aleu-rolite	2,34	14,0	I,23	—	I,55
"	2085		2,29	17,5	I,04	—	I,49
"	2087	"	2,23	16,9	I,13	—	I,49
"	2150	"	2,33	13,3	I,08	—	I,39
"	2222	"	2,30	15,7	I,30	—	I,45
"	2524	"	2,39	13,1	I,13	—	I,39
"	2555	"	2,38	15,2	I,00	—	I,24
"	2555	"	2,27	15,2	I,03	—	I,36
Borehole 74	2075	"	2,31	13,4	I,08	—	I,48
"	2055	"	2,23	15,3	I,19	—	I,43
"	2121	"	2,30	15,3	I,25	—	I,49
"	2165	"	2,75	5,2	I,20	—	I,21
"	2285	"	2,64	4,0	I,36	—	I,55
Borehole 62	2138	"	2,32	16,1	I,03	—	I,34
"	2170	"	2,56	17,6	I,16	—	I,55
"	2240	"	2,40	15,7	I,II	—	I,44
	2087						
Borehole 76	2406	"	2,17	20,7	0,85	—	I,29
"	2540	"	2,79	15,5	I,I3	—	I,40
"	2560	"	2,43	9,8	I,I3	—	I,36
"	2580	"	2,58	19,9	I,25	—	I,54
Borehole 66	2132	"	2,08	22,4	0,95	—	I,34
"	2185	"	2,00	21,0	I,07	—	I,45
Borehole 71	1970	"	2,26	15,2	I,12	—	I,23
"	2130	"	2,04	23,5	0,93	—	I,66
Borehole 116	2370	"	2,26	16,6	I,16	—	I,48
Borehole 69	2090	argil-lite	2,30	14,8	I,12	—	I,45
"	2090	"	2,30	14,8	I,12	—	I,45
Borehole 500	2111	"	2,25	17,8	I,21	—	I,39
"	2115	"	2,30	15,1	I,08	—	I,44
Borehole 234	2061	"	2,29	18,4	I,14	—	I,45
"	2110	"	2,30	16,8	I,21	—	I,42
"	2125	"	2,27	15,8	I,II	—	I,39
"	2150	"	2,25	18,5	I,06	—	I,48
"	2215	"	2,24	17,9	0,91	—	I,36
"	2120	"	2,31	15,5	I,07	—	I,39
"	2225	"	2,29	15,5	I,05	—	I,34

Commas represent decimal points.

Table 8. Continued

Sampling site (region, borehole)	Sampling interval, m	Rock	Density γ, g/cm³	Open Porosity, %	Heat conductivity of an absolutely dry sample W/m·deg.	Heat conductivity of sample in an air-dry state, W/m·degree	Heat conductivity of water-saturated sample, W/m·degree
Borehole 234	2255	argil-lite	2,33	I4,9	1,08	–	I,36
"	230I		2,39	I3,6	I,27	–	I,5I
"	2505	"	2,43	II,I	I,II	–	I,30
"	2505	"	2,38	I3,2	I,I4	–	I,35
"	25I9	"	2,4I	I4,2	0,98	–	I,35
"	2543	"	2,I0	20,2	0,99	–	I,2I
"	2758	"	2,45	9,66	I,24	–	I,45
"	2800	"	2,22	II,9	0,86	–	I,34
"	2800	"	2,43	10,5	I,I9	–	I,49
"	2860	"	2,8I	II,I	I,30	–	I,39
Borehole 74	207I	"	2,25	I7,4	I,08	–	–
"	2075	"	2,3I	I6,5	I,24	–	I,43
"	2075	"	2,27	I7,2	I,I3	–	I,49
"	2090	"	2,27	I6,0	0,99		I,00
"	2095	"	2,22	I8,9	I,I9	–	I,43
"	2I20	"	2,29	I6,5	I,II	–	I,36
"	2I30	"	2,24	I8,0	I,08	–	I,40
"	2235	"	2,30	I6,4	I,08	–	I,29
"	2270	"	2,26	I6,9	I,03	–	I,44
"	2285	"	2,3I	I6,2	I,05	–	I,34
"	2290	"	2,34	I7,2	I,0I	–	I,44
Borehole 52	2I40	"	–	–	I,I3	–	–
"	2I75	"	2,32	I9,9	I,II	–	I,40
"	22I5	"	2,33	I5,7	I,II	–	I,30
Borehole 76	2466	"	2,45	I0,2	I,2I	–	I,40
"	2470	"	2,47	9,5	I,34	–	I,5I
"	2470	"	2,42	II,I	I,2I	–	I,5I
"	2473	"	2,47	I0,6	I,27	–	I,44
"	2484	"	2,52	I2,8	I,I2	–	I,44
"	2495	"	2,45	II,7	0,95	–	I,44
Borehole 510	2095	"	2,00	25,I	0,90	–	I,08

Commas represent decimal points.

This can evidently be attributed to the fact that for sandstones, water is a wetting fluid and even an insignificant moistening leads to a considerable improvement in the contact between grains and therefore improves the heat conductivity of the rock. Similar results have also been obtained by some other researchers (B.A. Yakovlev, S.P. Vlasova, oral communication).

After comparing the heat conductivity in absolutely dry and completely water-saturated states, we conclude that the heat conductivity increments can be significant, in some cases attaining 90% or more. However, it must be noted that the moisture content correction to heat conductivity in an air-dry state is less. In our case (Kamchatkan samples) it exceeds 20% for only some samples.
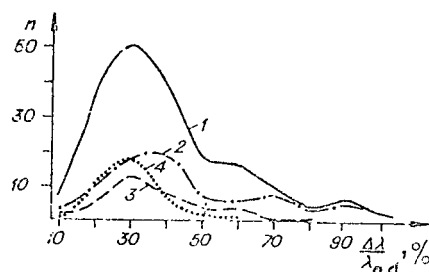
Figure 9. Variation curves of corrections for heat conductivity coefficient with moistening of rocks.

1 - sedimentary rocks of West Siberian Lowland and Kamchatkan Peninsula; 2 - sandstones; 3 - aleurolites; 4 - argillites.

Figure 9 shows the $\Delta\lambda/\lambda_{a.d.}$ variation curves /43 ($\Delta\lambda$ is the difference between the heat conductivities of dry and water-saturated samples; $\lambda_{a.d.}$ is the heat conductivity of an absolutely dry sample) for each type of rock (sandstones, aleurolites, argillites) and a general curve for all rocks. On each of these curves there is a clearly expressed peak which on the x-axis corresponds to a 30% increment in this ratio. Only for sandstones is it broader and falls between 30 and 40%. Most samples of the investigated sedimentary rocks obviously have an increment to heat conductivity in an absolutely dry state close to 30% as a result of water moistening (to total saturation). Accordingly, in large-scale determinations of the heat conductivity for such rocks the moisture content correction of 30% must be considered most probable.

The results of determination of the thermal properties of igneous and sedimentary rocks made at room temperature and atmospheric pressure on a large number of samples give some idea concerning the thermal parameters of the most characteristic rocks in the investigated regions. Such information was earlier completely lacking for these regions.

These data on the above-mentioned thermophysical constants of rocks and their dependence on composition, porosity and moisture under normal conditions made it possible to proceed to a study of the thermal properties of rocks at high temperatures.

CHAPTER III

HEAT CONDUCTIVITY OF ROCKS AT HIGH TEMPERATURES

Until now information on the effect of temperature on the thermal properties /44
of rocks has been extremely limited. The results of experiments for study of the
effect of temperature on the heat conductivity of rocks are given only in studies
by F. Birch (Birch, et al., 1949), W. D. Kingery (Kingery, 1954), K. Kawada
(Kawada, 1964), U. I. Moiseyenko, Z. A. Solov'yeva, V. A. Kutolin (1965, 1966,
1967).

The experiments of F. Birch dealt with several types of igneous and sedi-
mentary rocks heated to a temperature of 400°, including granite heated to 500°.
During recent years K. Kawada has obtained data on the behavior of the heat con-
ductivity coefficient for igneous rocks (Figure 10) in a broader temperature range
(20-600°). W. Kingery made similar experiments with forsterite with heating to
1400°.

In our laboratory the effect of temperature on the heat conductivity co-
efficient was studied on samples of different types of igneous rocks.

Measurement Apparatus and Method

Measurements of the heat conductivity coefficient at a high temperature
were made by the stationary heat flux method and the "plate" method, since they
are the best developed and measurements can be made with a high accuracy.

The shortcoming of the method, the duration of the experiment, in this case /45
is a positive consideration, since the prolonged presence of a rock in a station-
ary regime in a way simulates the state of rocks at different depth levels. The
measurements were made using samples of different igneous rocks 80 mm in diameter
and 5-15 mm high.

The apparatus used in determining heat conductivity of rocks at a tempera-
ture of 1400° was constructed by the Experimental Plant Siberian Department USSR
Academy of Sciences (Figure 11). It was described in a study by V. V. Goncharov
and A. F. Kolechkova (1963). The apparatus consists of two main parts: a furnace
for unilateral heating of the sample and a water-jet central calorimeter with a
guard ring; these were protected against the heat flux by a water jacket. The
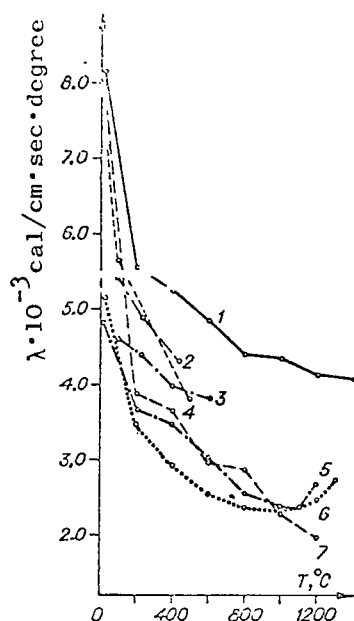
Figure 10. Dependence of heat conductivity coefficient of rocks on temperature, data from various authors.

1- olivinite; 2, 7- eclogite; 3, 5- diorite; 4, 6- granite; 5, 6, 7- from U. I. Moiseyenko, et al., (1965, 1966, 1967); 2, 3- from K. Kawada (1964); 4- from S. Clark and F. Birch (Birch, et al., 1949).

sample to be tested is placed on the calorimeter and guard ring. The electric heaters are carborundum or silicon carbide rods. The heat emitted by the electric heaters passes through the sample and is imparted to the water continuously flowing through the central calorimeter and guard ring. Thermometers (usually Beckmann thermometers) are /46 placed in the inlet and exit from the central calorimeter in the connecting pieces for measuring the temperature of the flowing water. The accuracy in determining the heat conductivity coefficient is 10%.

The dependence of heat conductivity on temperature was studied on samples of olivinite, granite, diorite, obsidian, eclogite, dolerite, and pyroxenitic gabbro.

Experimental Results

The experiments were begun with determination of the heat conductivity of olivinite; this made it possible to check the correctness of the method used because it was possible to compare the results with the data obtained

by W. Kingery for forsterite.

The olivinite sample was taken from a collection of rocks from the Monchegorsk (Kola Peninsula) region. Macroscopically, this is a dark gray rock, consisting of isometric olivine grains 1-2 mm in diameter. It could be seen under the microscope that the rock has a pandiomorphic texture and consists of entirely /47 of idiomorphic grains of completely fresh olivine and a small quantity of magnetite (Figure 12a); the olivine contains 12% fayalite molecules.

The sample exhibits the following changes after heating.[1] A dense, fine powder of reddish-brown hematite forms along the boundaries of the olivine grains

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1. V. A. Kutolin made the petrographic study of sections of rocks subjected to high temperatures.
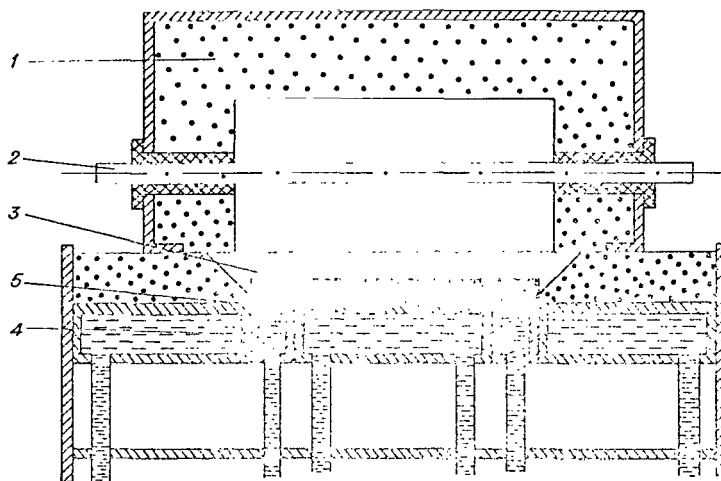
Figure 11. Apparatus for determining the heat conductivity
coefficient at high temperatures.

1 - furnace; 2 - heating element; 3 - sample; 4 - guard ring;
5 - calorimeter.

and along the tiny fissures within the grains in the section made from the lower
(cold) part of a sample heated to a temperature of 880° (see Figure 12b). It can
be clearly seen in the section along the plane transverse to the surface of
sample heating that with approach to the upper (hot) side of the sample there
is a displacement of the hematite powder by an equally fine magnetite powder
which is fully completed in the upper half of the sample. The magnetite powder
also appears within the olivine grains. Finally, a decomposition of the initial
olivine is observed in the section from the upper part of the sample. In the
early stage of this process the olivine grains manifest abundant inclusions of
fine magnetite grains, whereas the fine powder of this mineral disappears. The
boundaries between the olivine grains become less distinct. The more complete
decomposition to which entire mineral grains are sometimes subjected, and which
is manifested most clearly at the contact of several grains, involves a sub-
stitution of olivine by an aggregate of skeletal magnetite grains and fine
prisms of greenish orthopyroxene having direct extention. In many cases olivine
remnants persist amidst this aggregate in the form of simultaneously extinguish-
ing fragments (see Figure 12c). Thus, the observed decomposition of olivine
into magnetite and orthopyroxene with its oxidation by atmospheric oxygen is
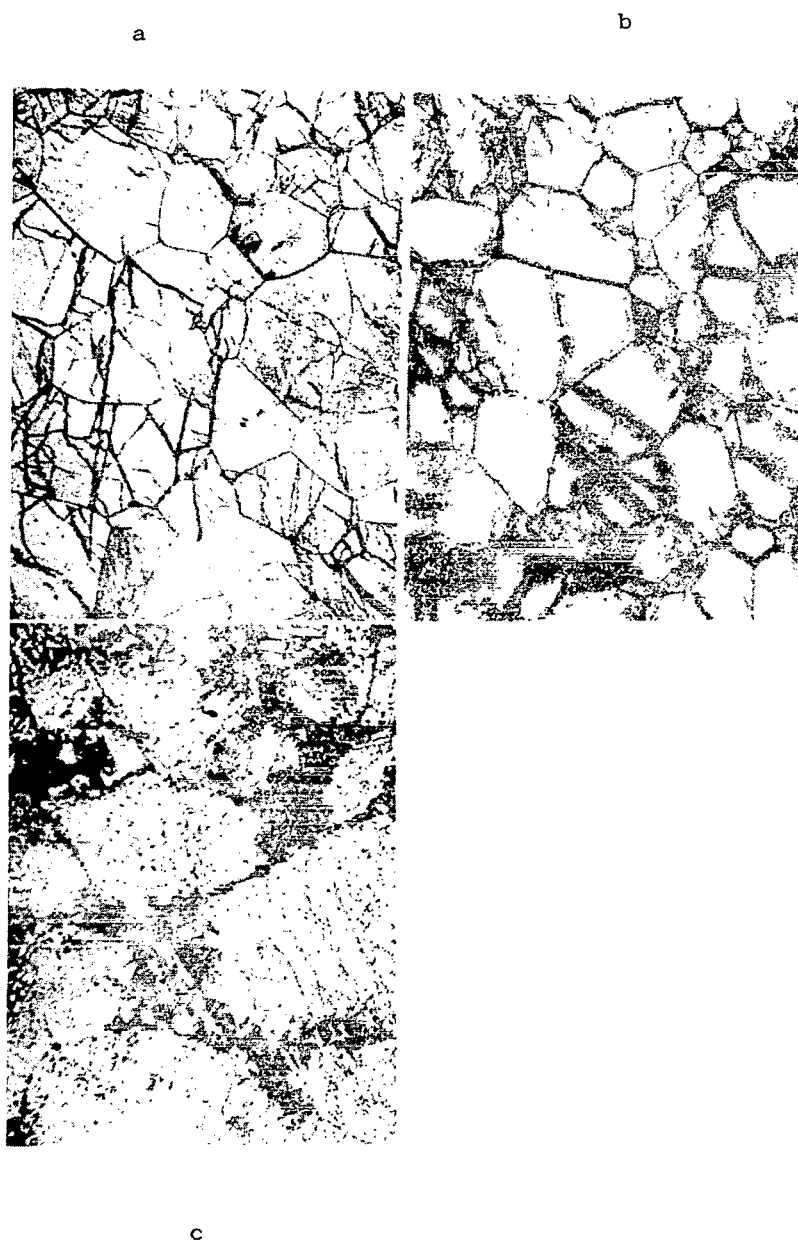similar to that described by I. D. Muir, et al. (Muir, et al., 1957) for the

41

a   b

c

Figure 12.   a - initial olivinite; b - olivinite from cold side;
             c - olivinite from hot side. Nicols. Magnification 21.6 X

metamorphized picritic basalts of the Hawaiian Islands, whose mechanism was discussed in detail by H. S. Joder and C. E. Tilly (1965).

The description presented above shows that during the sampling heating process the olivine decomposes; this evidently somewhat distorts the numerical heat conductivity characteristics. It is possible that they would be different if the olivine was heated in an inert medium. Nevertheless, it can be asserted that the order of magnitude of the numerical data and the general shape of the curve in change of heat conductivity with an increase in temperature do not exhibit significant changes because the intensive decomposition of olivine with the formation of orthopyroxene and magnetite occurred only in the upper part of the sample.

Figure 13 shows heat conductivity curves for olivinite based on our data and material tested by W. Kingery with heating to 1400°. The heat conductivity coefficient for olivinite decreases with a temperature increase. At room temperature it is 3.43 W/m·degree, whereas at a temperature of 1400° it is 1.47 W/m·degree. These heat conductivity values for olivinite are close to those obtained by W. Kingery for forsterite (see Figure 13); in our opinion this confirms the correctness of the selected method and the reliability of the results.

Figure 13. Dependence of heat conductivity coefficient of olivinite (1) and forsterite (2) on temperature.

The investigations were continued with similar experiments with leucocratic biotitic granite from the Belokurikhinskiy complex in the Altay. The measurements were made at temperatures 200, 400, 700, 1000, 1100, 1200, and 1300°. With a temperature increase the granite heat conductivity coefficient constantly decreases; changes of a particularly significant magnitude were observed to 500°; at temperatures above 1000° the heat conductivity increased. The experiments made with a series of samples duplicated the shape of the curve with a minimum at 1000°. Up to a temperature of 500° curve 6 (see Figure 10) coincides with the data published by F. Birch and S. Clark for this range. At higher temperatures it agrees well with the computations of K. Clark (Clark, 1956), A. Lawson (Lawson, 1958) and V. N. Zharkov (1958).

In all samples whose lower surfaces were heated to 770° and whose upper
surfaces were heated to 1300° there are two well-defined zones (Figure 14). A
lower zone, equal to approximately half the sample thickness, has a brick color.
It can be seen in the sections prepared from this part of the samples that in the
potash feldspar of granites there are fine brownish decomposition products which are
responsible for this color. Quartz and feldspars frequently have a nonuniform
undulating extention, revealing a considerable similarity to minerals from cata-
clastic rocks. The greatest change in exhibited by biotite; it first becomes   /50
brown and then, in the upper, hotter part of the zone, is packed with fine non-
translucent iron oxides.

The upper zone of gray samples includes glass which is readily distin-
guishable even macroscopically due to the thickness of the zone. The upper part
of the samples is covered by a thin lustrous encrustation of glass with remnants
of still unfused crystals. In the sections prepared from the lower part of the
upper zone there are distinct traces of fusion: the rock contains much trans-
parent colorless glass of acidic composition, amidst which there are individual
corroded grains and aggregates of grains of quartz, feldspar and ore mineral. It
can be seen in the sections from the upper glassy encrustation that the feld-
spars are completely fused and the rock has been transformed into a glass con-

Figure 14. Granite section. Lower part of figure, with heating
            to 770°; upper part of figure, with heating to 1300°.

taining highly corroded quartz grains and ore mineral; around the latter the glass has a brownish color.

The fusing of those samples of granite whose upper part was heated to 1300° and whose lower part was heated to 770° is propagated to the middle of the sample. If it is assumed that the temperature varies linearly from the upper to the lower surface of the sample, then its middle part must have a temperature of about 1000°. Thus, in our experiments the fusing of granites began /51 at a temperature of $\sim$ 1000°; this agrees well with data from other researchers on the fusing of granite in dry systems (Lebedev, 1964). A petrographic study makes it possible to explain the peculiarities of the curve of the dependence of the heat conductivity of granites on temperature (see Figure 10). The minimum on the curve corresponds precisely to the mean temperature, 1000°, that is, to the fusing point for granites.

Thus, with heating of granites to their fusing point their heat conductivity decreases, whereas after the onset of fusion it increases with a temperature increase.

We continued investigations of the heat conductivity of rocks at a high temperature in a series of experiments with obsidian (Caucasus). The rock consists of glass of an acidic composition in which one notes the fusion of very small microliths of feldspar, extremely nonuniformly scattered in the vitreous matrix, first oriented in subparallel fashion and collected into irregular bands creating a flow texture, and then exhibiting a felted aggregate. In very rare cases the rock contains phenocrysts of acidic plagioclase and biotite. During the heating process the upper half of the sample was fused, bulged, and acquired a slaggy texture with numerous large (0.5 to 1.0 cm) voids (Figure 15).

Figure 16 shows a curve of change in heat conductivity for obsidian with an increase in temperature. The heat conductivity of obsidian, in contrast to granite, increases with a temperature rise, attaining a maximum at 900°; then it decreases, possibly because of bubbling of the liquid melt. In gross chemical composition granite and obsidian are extremely close to one another, but they exhibit a diametrically opposite change in heat conductivity with a temperature increase, evidently the result of a difference in the structure of these rocks: granite is a crystalline rock whereas obsidian is an amorphous,

vitreous substance.

Determinations of the heat conductivity of rocks at high temperatures were also made on samples of basic rocks: eclogite, dolerite and pyroxenitic gabbro. We studied samples of eclogite from Kazakhstan, dolerite from the Berikul'skiy region of the Kuznetskiy Alatau and gabbro from the Eastern Sayan.

The initial eclogite has a porphyroblastic texture and consists of large, rounded idioblasts of rosy garnet enclosed in a nematoblastic matrix of pyroxene, blue-green alkaline amphibole, and quartz. In addition, in the section one notes /52 abundant fine grains of rutile and individual flakes of colorless mica. The garnet idioblasts usually contain numerous epidote inclusions.

The heat conductivity of eclogite decreases sharply with heating (Figure 17). The most significant decrease in heat conductivity is observed on the curve segment corresponding to temperatures from room temperature to 200°. In the first case the heat conductivity coefficient is 3.35 W/m·degree. At a tem-
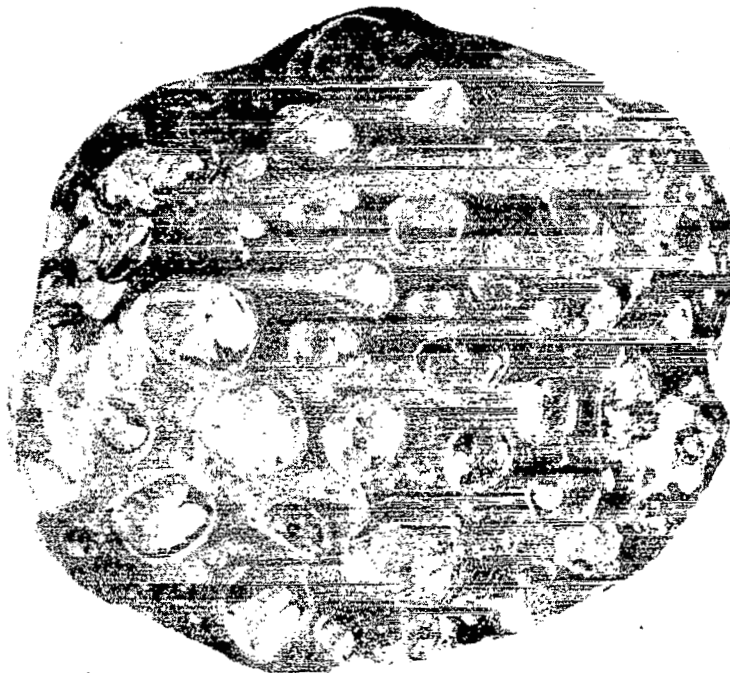

Figure 15. Obsidian after heating to 1000°C.

perature of 400° its value decreases to 1.53 W/m·degree, at 800° to 1.2 W/m·
degree and at a maximum temperature (1200°) to 0.82 W/m·degree, that is, de-
creases by a factor of four in comparison with the initial level. The rock
changes during heating as follows. Amphibole in sections from the lower part
of the sample (670°) changes from blue-green to brown-green, with a strong
pleochroism from green to brown, whereas other minerals do not change. In the
sections from the middle part of the sample (950°) the amphibole and garnet
are completely packed with opaque fine black iron oxides and the pyroxene is
completely replaced by a dark-grained aggregate of grayish-brownish-greenish
mineral similar to bowlingite, whereas the quartz remains unmodified. Finally,
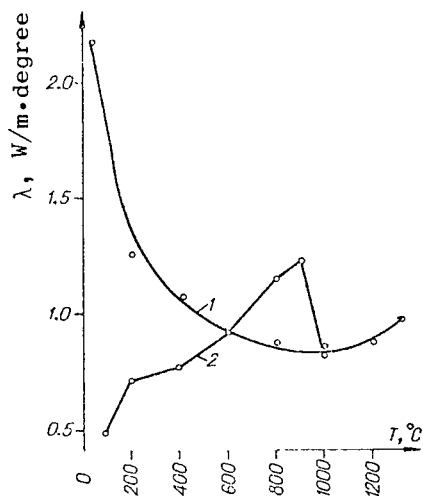the upper part of the sample, heated to 1200°,
constitutes a colorless or cinnamon-colored glass
with an abundance of skeletal grains of magnetite
and felted clusters of plagioclase microliths,
crystallized out during the slow cooling of the
sample, and also fused quartz grains. In addition,
in many cases there are sectors of uniform trans-
parent cinnamon-colored glass.

In measuring the heat conductivity of doler-
ite, the initial unmodified rock had a prismat-
ically granular structure and consisted of pris-
matic grains of plagioclase, between which there
are grains of a rosy-brownish titaniferous clino-

Figure 16. Dependence of
heat conductivity of granite
(1) and obsidian (2) on
temperature.

pyroxene, usually entirely substituted by highly
pleochroic brownish-green amphibole, as well as
finer rounded olivine grains. In addition, the
rock contains abundant flakes of a reddish-brown, highly pleochroic biotite and
fine granules of ore minerals. A small quantity of acidic untwinned plagioclase
can sometimes be observed in the interstices.

With heating of the dolerite to 500° the rock changes are manifested in an
intensification of amphibole pleochroism from greenish-brown to dark brown, al-
most black. In addition, an opacitized edge appeared around the grains. With
heating to 1100° the amphibole and olivine were completely opacitized and an
opacitization of the pyroxene began, although a considerable part of it remained

unmodified. Plagioclase exhibited no changes.

In Figure 17 the changes in the heat conductivity of dolerite with heating to 1100° are represented by curve 3. The heat conductivity of dolerite at room temperature is lower than for eclogite, 1.57 W/m·degree. The nature of the change in heat conductivity for dolerite is different than for eclogite. At the beginning of heating the heat conductivity of dolerite decreases and at a temperature of 200° it attains 1.26 W/m·degree. With a further temperature increase λ increases monotonically to 1.29 W/m· degree at 600°, attaining a maximum at 800° (1.43 W/m·degree). This is followed by a heat conductivity decrease and at 1100° it is already 1.31 W/m· degree.

Figure 17 shows the curve of change for λ for dolerite and pyroxene gabbro.[2] These curves have an identical shape although the peaks are displaced relative to one another. In the case of pyroxene gabbro the heat conductivity decrease with heating to 200° is also replaced by its increase with a peak at 600°. A further decrease in its heat conductivity is greater than for dolerite. Its value decreases from 1.49 W/m·degree at 600° to 1.1 W/m·degree at 1100°.
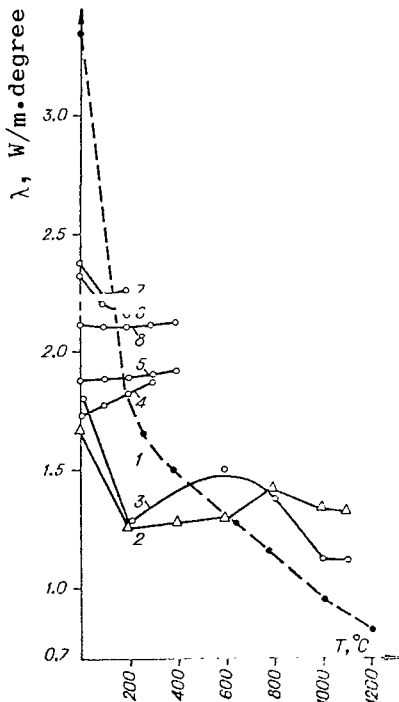

Figure 17. Dependence of heat conductivity of eclogite (1), pyroxene gabbro (2), dolerite (3), anorthosite (4), gabbro (5, 7) and diabase (6, 8) on temperature; 4-8) data from S. Clark and F. Birch (Birch, et al., 1949).

A comparison of the described heat conductivity curves for dolerite and pyroxene gabbro with the data given by F. Birch on the heat conductivity of gabbro, anorthosite and diabase in the temperature range 0-300° (see Figure 18) shows that the heat conductivity of anorthosite increases continuously to 300° (curve 4). At the same time, the heat conductivity of gabbro in this range in one case (curve 5) increases (to be sure, very insignificantly, by only 0.02 W/m·degree), whereas in another case (curve 7) it decreases with a temperature increase. A similar

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2. Pyroxene gabbro is close in structure to dolerite and its petrographic description is not given for that reason.

picture is observed for diabase. In one case (curve 8) a decrease in heat conductivity, being 0.11 W/m·degree per 100°, is replaced by its increase by 0.02 W/m·degree with heating to 200°; in another case (curve 6) the heat conductivity of diabase also decreases, but this decrease, like the subsequent increase, is extremely insignificant.

It was established in these experiments that the nature of the change in the coefficient λ is not dependent on the rock basicity. In actuality, the heat conductivity of such rocks as granite, olivinite and eclogite decreases continuously with a temperature increase, whereas for dolerite, gabbro, and possibly diabase it decreases, but not constantly. There is some increase in λ from 200 to 800°.

In obsidian and anorthosite the heat conductivity coefficient increases with heating (in anorthosite at least to 300°; no experimental data are available for higher temperatures). The peculiarity of the change in heat conductivity for gabbro and dolerite may evidently be related to the presence of basic plagioclase in their composition, since according to data published by F. Birch, anorthosite is characterized by an increase in heat conductivity with a temperature increase. No plagioclase is present in rocks of the first group. It is interesting that the denser isochemical equivalent of rocks of basic composition, eclogite, differs from the latter in the nature of heat conductivity change with a temperature increase.

Table 9 gives experimental data for the investigated rocks for determining their heat conductivity λ at a high temperature.

The data in Table 9 on λ for some igneous rocks at different temperatures are important for computing the heat flow and deep temperatures in the earth's crust.

Table 10 gives the temperature corrections for heat conductivity which must be taken into account in computing the heat flow.

The data in Tables 9 and 10 were used in computing the temperature in the /57 earth's crust to the Mohorovicic discontinuity for some places on the West Siberian Lowland. The collected data on the thermal properties of some rocks at high temperatures also help in understanding the phenomena transpiring in the earth's crust: the mechanism of heat accumulation, the nature of local

TABLE 9. HEAT CONDUCTIVITY COEFFICIENT FOR DIFFERENT ROCKS
AT HIGH TEMPERATURES

Temperature, °C	Olivinite, W/m·degree	Leucocratic granite, W/m·degree	Obsidian, W/m·degree	Eclogite, W/m·degree	Dolerite, W/m·degree	Pyroxene gabbro, W/m·degree
20	3,43	2,16	-	3,35	I,57	I,80
200	2,53	I,45	0,90	I,24	0,85	I,27
400	2,20	I,23	I,08	I,2I	0,88	I,40
600	2,03	I,06	I,2I	I,20	0,92	I,49
800	I,84	0,99	I,76	I,09	I,10	I,37
1000	I,83	0,97	I,7I	0,95	0,97	I,07
1100	-	-	I,3I	-	0,95	I,00
1200	I,73	I,03	-	0,78	-	-
1300	-	I,I4	-	-	-	-
1400	I,7I	-	-	-	-	-

Commas represent decimal points.

TABLE 10. CORRECTIONS TO THE HEAT CONDUCTIVITY COEFFICIENT
FOR ROCK WITH TEMPERATURE CHANGE

Rock	Change in heat conductivity per 10°, %	Reference
Limestone	0.9-2.4	Birch, et al., 1949
"	3.0	Roy, 1963
Schist	0.9	Birch, et al., 1949
Chalk	2.0	Roy, 1963
Quartzy sandstone	2.2	Birch, et al., 1949
Marble	1.9-2.1	"
Diabase	0.0-015	"
Basalt	1.2	Kawada, 1964
Diorite	0.9-1.0	"
"		Moiseyenko, Kutolin, 1966
Granite	0.7-1.4	Birch, et al., 1949
Granite	1.9	Moiseyenko, et al., 1965
Pyroxene gabbro	1.7	Moiseyenko, et al., 1967
Olivinite	2.6	" 1965
Eclogite	3.0	" 1967

Note: Since the minimum depth of boreholes was 3000 to 4000 M, the temperature interval for computing corrections was in accordance with these depths.

sources of magma formation, metamorphism and other deep processes.

Mechanism of Rock Heat Conductivity

The heat conductivity of monolithic igneous rocks with insignificant porosity can be represented in the form

$$\lambda = \lambda_{el} + \lambda_{lat},$$

where

λ_{el} is the heat conductivity component caused by the heat transfer by electrons and holes, and

λ_{lat} is the crystal lattice.

In the temperature range 20 to 1400° at which our experiments were made, the heat conductivity exciton mechanism evidently plays no role (B'yub), 1962).

Electron heat conductivity. At temperatures above room temperature λ_{el} can be written in the following form (Drubble, Goldschmidt, 1963):

$$\lambda_{el} = 1.4 \cdot 10^{-8} T \left[(\sigma_n + \sigma_{hole}) + \frac{1}{2} \cdot \frac{\sigma_n \cdot \sigma_{hole}}{\sigma_n + \sigma_{hole}} \left(4 + \frac{E_G}{KT} \right)^2 \right] \quad (1)$$

Here σ_n and σ_{hole} are the electron conductivities caused by electrons and holes /58 respectively; E_G is the width of the forbidden zone. Using this expression λ_{el} can be estimated. The second term in the cited formula will obviously be maximum when $\sigma_n = \sigma_{hole}$. We therefore assume that $\sigma_n = \sigma_{hole} = 1/2 \sigma$. Then expression (1) can be rewritten:

$$\lambda_{el} = 1.4 \cdot 10^{-8} \sigma T \left[1 + \frac{1}{8} \left(4 + \frac{E_G}{KT} \right)^2 \right]. \quad (2)$$

In (2) we assume that $E_G = 7$ eV (7 eV is the width of the forbidden zone for Al_2O_3, being one of the principal rock components. This compound has the broadest forbidden zone among minerals). According to our experimental data (see Table 1), at a temperature of about 300°K, $\sigma = 10^{-12} - 10^{-9}$ ohm$^{-1} \cdot$cm^{-1}, and at a temperature of about 1470°, $10^{-4} - 10^{-3}$ ohm$^{-1} \cdot$cm^{-1}. After substituting these values into formula (2), we obtain:

$$\lambda_{el} \ (300°K) \lesssim 10^{-10} \ W/cm \cdot degree;$$
$$\lambda_{el} \ (1470°K) \lesssim 10^{-5} \ W/cm \cdot degree$$

$$(3)$$

It should be noted that for igneous rocks the λ_{el} values in (3) are highest at the temperatures given in parentheses.

Laboratory measurements in the entire indicated temperature range for the total heat conductivity of igneous rocks give about 10^{-2} W/cm·degree (see Table 5). Thus, electron heat conductivity is an insignificantly small part of the total heat conductivity for the examined rocks. Lattice heat conductivity is most important.

<u>Lattice heat conductivity</u>. The problem of crystal lattice heat transfer is usually reduced to a study of the motion of phonons in a potential field created by the medium crystal lattice. Accordingly, λ_{lat} is dependent on the rock structure. The rocks which we studied can be divided into two groups: amorphous and polycrystalline. The first group includes obsidian, the second group includes granite, olivinite, gabbro, and others.

In the case of rocks with an amorphous structure the latter is almost /59 entirely disordered. The heat transfer process conforms to the theory of random processes; at the considered temperatures this gives the dependence $\lambda \sim T$ (Zayman, 1962). Our data (see Figure 16, curve 2) apparently correspond to the prediction of the theory of a linear increase in λ with a temperature increase. However, at temperatures greater than 900° the heat conductivity decreases. This is probably attributable to an intensive release of volatile components in a molten state, as indicated by the appearance of the sample after the experiment was terminated (see Figure 15).

In polycrystalline structures the heat conductivity is determined both by the scattering of phonons on crystalline grains and their scattering on one another due to anharmonicity. At temperatures above the Debye temperature θ the scattering of phonons on one another as a result of anharmonicity leads to the dependence (Zayman, 1962)

$$\lambda = \lambda_0 \frac{1}{T} ,$$

which is qualitatively confirmed by our curves (see Figure 10).

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SUMMARY

Even at this stage in the investigation the study of the electric and /60
thermal properties of rocks made it possible to determine the distinctive
characteristics of their behavior under different thermodynamic conditions.

Temperature exerts a strong effect on rock resistivity. In the studied
temperature range (20-1200°), which can correspond to depths of 80-100 km, it
varies by several orders of magnitude. It should be noted that the difference
in resistivities for rocks of different compositions decreases at high tempera-
tures. However, when pressure is superposed, this smoothing effect can evi-
dently change. Accordingly, another objective of our investigations was a
study of the effect of pressure on rock resistivity.

Under the influence of unilateral pressure at room temperature a complex
change in resistivity is observed. It first decreases and then increases. The
resistivity minimum is observed in different rocks at different pressures. Under
unilateral pressure there is an interrelationship between resistivity, volu-
metric weight, and porosity of rocks and their deformations. However, predom-
inantly unilateral pressure (one-sided up to 20,000 kg/cm^3, but hydrostatic
pressure not above 2000-3000 kg/cm^3) and room temperature does not correspond
to conditions at great depths in the earth's crust and mantle. Accordingly,
the determined dependences can be used for the most part in studying rock de-
formations.

The results of experiments for determining rock resistivity under the
simultaneous influence of temperature and pressure are most valuable. In /61
these experiments a temperature of 500° and a pressure of 30 kbar were attained.
Rock resistivity decreases with a pressure and temperature increase.

Analysis of the collected experimental data made it possible to draw con-
clusions concerning the conductivity mechanism in rocks. Formulation of speci-
fic experiments for determining the concentration and sign of charge carriers
is required for drawing more rigorous conclusions.

On the one hand the study of the thermal properties of rocks both under
ordinary conditions and at high temperatures made it possible to obtain the
first information concerning the thermal constants of the most widely occurring

rocks in the investigated regions; on the other hand, it made it possible to clarify the dependence of heat conductivity on temperature. It was found that the nature of the change in heat conductivity with a temperature increase is dependent on structure. In amorphous rocks (obsidian) the heat conductivity increases with a temperature increase, whereas in crystalline rocks (most of the igneous rocks) it decreases.

The results of these experiments were used in computing the temperatures of seismic discontinuities along deep seismic sounding profiles on the West Siberian Lowland.

The observed decrease in heat conductivity with a temperature increase helps us to understand the nature of local centers of magma formation in the earth's crust and upper mantle at depths less than might be expected on the basis of the geothermal gradient. Reliable interpretation of geophysical data, superdeep drilling, and study of deep processes are impossible without taking into account the patterns of change in the physical properties of rocks at high temperatures and pressures; this makes investigations in this direction particularly important.

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